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PHYSICAL SUBJECTS

ATMOSPHERIC
ELECTRICITY

B. F. J. SCHONLAND

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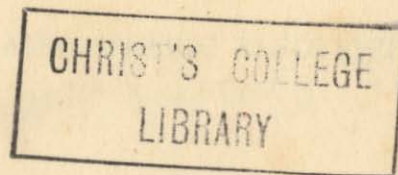
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WITH 20 DIAGRAMS

TO

C. T. R. WILSON

CONTENTS

CHAPTER I

THE IONISATION OF THE ATMOSPHERE

THE electrical conductivity of the air, first established by Linss in the year 1887, plays an important part in all the phenomena included under the title Atmospheric Electricity. Since 1901, when Elster and Geitel and C. T. R. Wilson independently discovered that this conductivity is due to the presence of small carriers of positive and negative charge called ions, the nature and the origin of the ions have been the subject of many investigations. Two different paths have been followed, the one dealing with the number and the nature of the atmospheric ions, the other with the radiations producing them. The two paths converge on the problem of the ionisation-balance of the atmosphere, the manner in which an equilibrium is maintained between the rate of creation and the rate of disappearance of the ions. It is only of recent years that an answer to this problem has emerged.

1. The Atmospheric Ions, Small and Large. A gas molecule is ionised when some external agency supplies the energy necessary to remove an electron from it, whereupon the positively charged residue of the molecule and the ejected electron, both of which under ordinary pressures very soon attach themselves to one or more uncharged molecules, form what is called an ion-pair. Such ions are known as "normal", "small", or "fast", and when found in the atmosphere have mobilities of the same order as those determined for pure dry air in laboratory experiments, about 1·5 cm./sec. in a field of gradient 1 v./cm.

The presence in the air of minute particles of dust and salt, of the products of combustion, and of extremely small drops of water even when the air is unsaturated, leads very often to the capture of a normal ion by one of these condensation or Aitken nuclei, as they are termed, and so to the formation of another type of ion of low mobility,

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ranging from 0·008 to 0·0003 cm./sec., known as " large ", " slow ", or " Langevin " ions. Roughly speaking, one-third of the condensation nuclei present in the air is electrically neutral, the other two-thirds carry, nearly equally, positive and negative charges captured from small ions. The interplay between the large and the small ions and the uncharged nuclei is of considerable importance.

Besides these two main types of atmospheric ion, a group of " intermediate " ions whose mobilities lie between 0·2 and 0·01 cm./sec. has been found to be present under special meteorological conditions such as low relative humidity. Some of these are known to be associated with molecular clusters of sulphuric acid from industrial processes.[1]

It has been found that the great majority of the ions, large and small, carry a single elementary charge, either positive or negative.[2] Over land areas the large ions considerably outnumber the small ones, often in the ratio of ten to one. Thus while the numbers, n_+ and n_-, of small ions of each sign per c.c. of the air over land usually range from 300 to 1000, those of the large ions, N_+ and N_-, amount to from 1000 to 80,000 per c.c. Since practically all the large ions are formed from the capture of small ones by nuclei, an increase in the number of these nuclei, such as is due to the fires of an industrial area, will increase the proportion of the large ions in the air at the expense of the small ions.

Over the oceans this state of affairs is reversed ; the air away from land areas is but poorly supplied with nuclei and the large ions are in the minority. The mean values of n_+ and n_- at sea lie between 500 and 700 per c.c., while N_+ and N_- are about 200 per c.c. In view of their comparative freedom from capture by nuclei, it is at first sight rather surprising that the number of small ions per c.c. in mid-ocean differs little from the same quantity over the land, where the average values of n_+ and n_- are also about 600 per c.c. This is, however, to be explained (§ 12) by the fact that the decreased rate of disappearance

of the small ions over the sea is balanced by a decrease in the rate at which they are produced.

2. The Determination of the Number of Ions per c.c. To determine the numbers of the small ions, positive and negative, per c.c. (n_+ and n_-) an arrangement known as an Ebert ion-counter is usually employed. This consists in essentials of a long earthed metal tube A (Fig. 1) on the axis of which is mounted a charged insulated rod B, connected to a quartz fibre electroscope. Air is sucked through the tube for about five minutes by means of a clock-work driven fan F and the speed of the air-stream is arranged to be slow enough for all the small ions entering A with charges opposite in sign to that on the central rod to be drawn to B before reaching the bottom of the tube. The speed of flow must at the same time be great enough to make it impossible for any appreciable number of
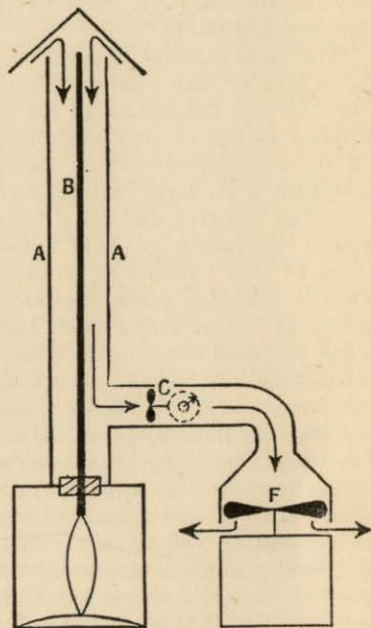

FIG. 1.—Ebert ion-counter.

large ions to have time to reach the rod. If v is the volume of the air sucked through in the course of the measurement (indicated by an anemometer C registering its revolutions on a dial), the quantity of electricity gathered in by the central electroscope system is ven_-, where e is the elementary charge on a single ion and the rod

is supposed to have been positively charged. If then C is the capacity of the whole central system, rod and electroscope, and V and V' are the potentials indicated by the electroscope before and after the experiment,

$$C(V - V') = ven_-,$$

an equation from which n_- can be found. To determine n_+, the measurement must be repeated with the rod B charged negatively.

The same method, using a longer tube, a stronger field in its interior, and a slower draught of air, and thus giving time for the capture of all the slow ions as well, can be applied to the determination of $N + n$. A combination of the two determinations then yields N_+ and N_-, the numbers of the large ions per c.c. of each sign.

The values found for n_+ and n_- are not usually the same; the ratio n_+/n_- over both land and sea has a mean value of 1·22. Although, as explained in § 22, the earth's electric field may sometimes affect the apparatus in such a way as to give a spuriously high ratio, Swann, who has made a special study of the point, concludes that when all precautions are taken a real excess of positive ions must be admitted. Probably the main factor causing this excess is the superior mobility (k) of the small negative ion; Nolan has found that k_-/k_+ has the value 1·16. A lower mobility implies a lower velocity of thermal agitation for the positive than for the negative ion, and consequently less likelihood of capture by condensation nuclei. This general excess of positive small ions does not necessarily involve a net positive charge in the air; it is easily compensated for by a comparatively small excess in the much larger number of large negative ions.

An instrument known as the Aitken dust-counter is often used to find the total number of nuclei, charged and uncharged, per c.c. It consists of a small expansion chamber filled with a sample of the air and saturated with water-vapour. The sudden movement of a piston gives an adiabatic expansion which causes drops of water to condense upon the nuclei; the resulting cloud falls

upon a glass plate divided into squares and the number of droplets is counted under a microscope. Though called a dust-counter it does not record the coarser dust in the air.[3]

In a recent improvement on the Aitken counter (Nolan [3]) a much larger sample of saturated air is expanded in a cylindrical vessel. The opacity of the cloud then gives a relative measurement of the concentration of nuclei.

3. The Polar Conductivities and the Ionic Mobilities.

A simple relation holds between the conductivity of the air and the numbers per c.c. and the mobilities of the ions responsible for conduction. Consider a point in the atmosphere at which the electric field intensity is F and the specific conductivity, the reciprocal of the specific resistance, is λ. The conduction current, i, through unit area drawn perpendicular to the field is then equal to $F\lambda$, for F is numerically equal to the potential difference, and $1/\lambda$ is the resistance, between the ends of a unit cube of the air. In this equation we assume that conditions are such that Ohm's Law can be applied to the air. Since the current is carried by positive and negative streams of ions moving in opposite directions we may write

$$i = F(\lambda_+ + \lambda_-), \quad . \qquad . \qquad . \quad (1.1)$$

where λ_+ and λ_- are called the polar conductivities.

Actually the existence of the two main types of ion, large and small, leads to the flow of four different ion-streams, two up and two down if the field is vertical, in each of which the velocity is different, being given by $F \times k$, where k is the corresponding mobility. The total current is thus

$$i = Fe(n_+k_+ + n_-k_- + N_+K_+ + N_-K_-), \quad (1.2)$$

where k_+ and k_-, K_+ and K_-, are the mobilities of the positive and the negative ions, large and small respectively, and e is the elementary charge. From (1.1) and (1.2), if the oppositely moving streams are separated,

$$\lambda_+ = e(k_+n_+ + K_+N_+) \text{ and } \lambda_- = e(k_-n_- + K_-N_-). \quad (1.3)$$

In clear country air over land, n_+ and n_- number about 600 per c.c., while N_+ and N_- are about 2000 per c.c.;

k_+ and k_- are 1·5 approximately and K_+ and K_-, 0·0004. Here, therefore, the ratio of the first to the second term in each bracket above is 1100 : 1 and most of the conductivity of the air is due to the small ions. Near large towns, however, n is usually of the order of 100 per c.c. and N, on account of the heavy pollution of the atmosphere may be 50,000. Under these conditions the second term would be one-seventh of the first and the large ions would play a part in the transfer of electricity through the air.*

Nolan and Nolan have obtained the polar conductivities by separate measurement of the n's and k's in eqn. (1.3) above. It is, however, more usual to find λ directly, from measurements of the rate which the charge on an insulated body exposed to the free air is dissipated by the flow of ions to it. Consider a negatively charged body exposed to moving air and let σ be the density of the electrification on any portion of area dS. Positive ions will move to the body, and since the field F just outside the surface is $4\pi\sigma$, the current per unit area tending to dissipate the charge will be $F\lambda_+ = 4\pi\sigma\lambda_+$. The total dissipation current, obtained by integrating over the whole exposed surface, is $4\pi\lambda_+ \iint \sigma dS$ or $4\pi\lambda_+ Q$, where Q is the charge exposed to the air. Since this current represents the rate at which the body losses charge, we have $-dQ/dt = 4\pi\lambda_+ Q$, a relation which provides a means of determining λ_+. A similar equation, with λ_- substituted for λ_+, holds for the case of a positively charged body.[4]

An arrangement used for this purpose is due to Gerdien (Fig. 2); it resembles the Ebert apparatus already described, but the air-stream is now made so rapid and the field in the interior of the tube so weak, that only a very small fraction of the total number of ions passing through A has time to reach the central rod. In this case the numbers of ions per c.c. are not appreciably altered by the experiment and Ohm's Law applies.

* The methods used to determine ionic mobilities are described in text-books on the conduction of electricity through gases.

If the potential V of the negatively charged central electroscope system is observed to rise at a rate dV/dt and if C is the total capacity of the central system, rod and electroscope, we have $-dQ/dt = -CdV/dt$. Also, if C' is the capacity of that portion of the central system exposed to the stream of air, the rod B and its support, $Q = C'V$, consequently

$$-CdV/dt = 4\pi\lambda_+ C'V, \qquad . \qquad . \qquad (1.4)$$

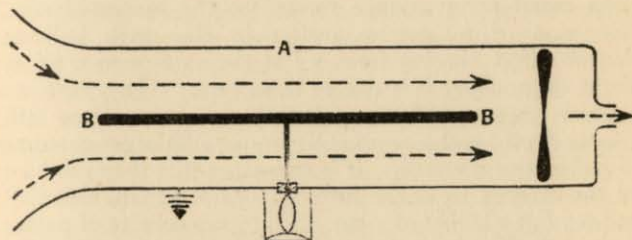a relation from which λ_+ can be determined. A similar

FIG. 2.—Gerdien conductivity apparatus.

experiment with the rod positively charged yields λ_-. The use made of the values found for the polar conductivities and an adaptation of the Gerdien method due to Schering are described in Chapter II, § 22.

4. The Ionising Agencies. Three principal agencies are responsible for the ionisation of the lower atmosphere : radiations from radioactive substances in the surface of the earth, radiations from radioactive matter present in the air itself, and the cosmic rays. For a discussion of their relative importance it is convenient to use the symbol q to represent the number of " pairs of ions produced per c.c. per second in air at N.T.P.", a quantity which will be referred to as the ionising power of the radiation. This phrase in inverted commas is usually represented by the symbol I. Thus the statement $q = 4I$ means that, at the point in question, the radiations create four pairs of ions per c.c. per second, if the medium there is air at N.T.P. The contributions of the three different agencies

just mentioned to the total ionising power, q, are distinguished by the symbols q_e, q_a, and q_h.

5. The Effect of Radioactive Matter in the Earth.
Uranium, thorium, and their products are widely distributed through the earth's crust, whose surface layers therefore emit α, β, and γ rays into the air. Since the α rays can emerge only from a very thin layer and can ionise only the first few cm. of the air, their effect is inconsiderable. The β rays can come from greater depths and penetrate much further into the atmosphere. Calculations based upon average values for the measured concentration of radioactive matter in the earth indicate that an effect ranging from $1\,I$ at the surface to $0\cdot1\,I$ at about 10 m. above it is due to these rays. The γ rays are a more important factor, for they can come from still greater depths and consequently from a still larger quantity of radioactive material. It is estimated that they produce on the average $3\,I$ at the surface, $1\cdot5\,I$ at 150 m. above it, and $0\cdot3\,I$ at a height of 1 km. The γ ray activity of potassium introduces another source of ionisation, for the potassium content of the earth's crust is considerable. No very detailed survey of this question has yet been made, but measurements indicate that the contribution of potassium may exceed that of the combined uranium-radium and thorium families. The average value for the ionising power of the earth-radiations near the ground has been found directly to lie between 2 and $10\,I$, which is in accord with the measurements of the radioactive content of the earth's crust. Local surface and geological conditions naturally have a marked effect upon the value of q_e at any given spot.

The radioactive content of sea-water is found to be very small in comparison with that of soil and rocks, and over the oceans the earth-radiation effect is negligible.

6. The Effect of Radioactive Matter in the Air.
The atmosphere contains a considerable quantity of the radioactive emanations, radon and thoron, and their successive products. A wire charged negatively to a potential of a few hundred volts and exposed to the air

collects an easily detectable quantity of radium A and thorium A by recoil from the disintegration of these emanations. The gases themselves arise from the decay of radium and thorium within the earth, from which they escape by diffusion, by thermal convection, and as a result of decreases in the external atmospheric pressure. The greater portion of the ionisation thus produced is due to the α rays from the emanations themselves and from their A and C products, for here, unlike the case of the radiations from the earth, there is no absorbing layer between the disintegrating atom and the air. The radioactive gases and their products are fairly evenly distributed by atmospheric turbulence through the lower air. This "air-radiation" effect can be calculated from radioactive data if the emanation content of the air is determined. The emanations may be removed from a measured volume of air by drawing it through tubes cooled with liquid air or filled with coconut charcoal. The quantity of emanation condensed or absorbed is then determined by the effect produced in an ionisation electroscope, corrected for the decay of the material during the time of the experiment. A more direct method is to fill a closed ionisation chamber first with ordinary air and then with air which has been freed by the above method from the emanations and from their products, and to determine the saturation currents in the two cases. If these are i_1 and i_2 respectively, v the volume of the vessel and e the elementary charge, the value of the air-radiation ionising power, q_a, is given by $i_1 - i_2 = q_a v e$.

The results of such measurements agree with estimates based upon observations of the average emanation content of the air in placing the air-radiation effect at about $2\,I$ in the neighbourhood of the earth's surface. Of this about 55 per cent. is due to radon and its products and the remainder to the thoron series. It varies somewhat with locality and with those meteorological conditions which determine the rate of escape of the emanations. Apart from the effect of strong upward winds in elevating the richer surface layers, the emanation content of the air

should decrease fairly rapidly with height. Direct information on this point is difficult to obtain, but what exists indicates that the air-radiation effect falls to less than 2 per cent. of its surface value at a height of 5 km.

The effect, like that of the earth-radiation, is extremely small over the oceans, where the emanation content of the air is only about 1 per cent. of its value over land areas.

7. The Effect of the Cosmic Rays. The third factor in the ionisation of the atmosphere is a radiation of peculiar interest, whose origin and nature are still the subject of much investigation. It has a penetrating power which is considerably greater than that of the most penetrating γ rays from radioactive bodies, and it travels to the earth's surface in a downward direction, undergoing a certain amount of absorption on the way. Its ionising power at sea-level over both land and ocean areas when measured in closed metal vessels varies from about $2 \cdot 0 \, I$ in low geomagnetic latitudes [6] to $1 \cdot 5 \, I$ near the magnetic equator, the earth's field preventing the slower ionising particles from reaching the equatorial belt.

Until the year 1911 it was generally thought that the ionisation inside a closed vessel was entirely due to radiations from the walls and from radioactive matter in the earth and the outside air. It was therefore to be expected that this ionisation would decrease when the vessel was raised above the surface of the earth. This was tested in balloon ascents by Hess in 1911, and Kolhörster in 1913, who found that the diminution with height was appreciable only during the first kilometre, after which an increase occurred. These observations led Hess to postulate an ionising radiation coming from above and suffering partial absorption in the atmosphere.

The ionisation due to this "cosmic" radiation (so-called because it is now known to originate outside the atmosphere and to have little connection with the sun) increases rapidly with height above sea-level, rising tenfold in the first 5 km.

8. The Ionisation Inside a Closed Vessel. It is a matter of considerable difficulty to determine separ-

ately the ionising powers of the earth-radiation and of the cosmic rays. Calculations of the former must be based upon measurements of the radioactive content of the earth's crust, and these are necessarily tedious and rough. To proceed further, one must examine the ionisation produced in the air inside a sealed metal chamber under conditions which are modified from those existing in the free air by the presence of the metal walls. This involves correction to free-air conditions. A further disadvantage is that a new ionising agency is introduced, the radioactive matter in and on the inside walls. The ionising power of this "wall-effect", or the "residual ionisation" of the vessel as it is called, is a constant depending on the material of which the walls are made and the cleanliness of the inner surface. It is usually represented by the symbol q_0.

The total ionising power of all the radiations operating upon the gas in a closed vessel may be called q' and the modified effects of the three main ionising agencies q'_e, q'_a and q'_h; we then have

$$q' = q'_e + q'_a + q'_h + q_0. \qquad . \qquad . \ (1.5)$$

We may suppose the vessel to take the form of a sealed ionisation chamber filled with air at N.T.P. which has been freed from radioactive matter. The saturation current carried by the ions to the central collecting electrode is then given by $i = q've$, where v is the volume of the vessel and e the elementary charge on an ion.

Various methods have been employed to separate out the four terms on the right-hand side of eqn. (1.5) from one another. The most satisfactory of these begins by determining q_0. The instrument is taken to a large snow-fed lake whose radioactive content, like that of the sea, is negligible. When it is sunk to the depth of a few feet in such a lake, the effects of earth and air-radiations are non-existent and eqn. (1.5) becomes $q' = q'_h + q_0$; on lowering the vessel still further, q'_h diminishes as the cosmic radiation becomes absorbed in the increasing thickness of water, and it is possible to reach depths at which the

saturation current is practically constant and unaffected by further sinking of the instrument. At this stage $q' = q_0$, and the residual ionisation is determined. By combining this observation with that made close to the surface of the lake, q'_h may be found.

Since the effects of the cosmic radiation and of the walls of the vessel are both independent of local conditions other than height above sea-level, they remain the same when the instrument is transported to a site on land at the same level. A measurement of the saturation current in the chamber under land conditions gives q' as stated by eqn. (1.5), and hence the sum of the earth and air effects, q'_e and q'_a can be found. In actual practice the air-radiation effect within a closed vessel is extremely small; the walls exclude α and β rays coming from outside and reduce q'_a to the ionising power of the γ rays emitted by the B and C products of the radio-active emanations. This amounts to but 4 per cent. of the effect produced by the air-radiation in the free atmosphere, in the case of a vessel with steel walls a few millimetres thick. It is therefore sufficient to determine the air-radiation effect, q'_a, by calculation based upon a knowledge of the emanation content of the air and of the nature of the walls of the vessel. The fourth factor, the effect of the earth-radiation, q'_e, is then obtained by substitution in eqn. (1.5).

To pass from these closed vessel determinations of the effect of the radioactive matter in the earth to the effect to be expected in the free air, experiments may be made with screens of different thicknesses placed around the ionisation chamber and the values thus obtained for the earth-radiation effect extrapolated to zero thickness. In doing so account has to be taken of the effect of the screens upon q'_h.

To derive the true ionising effect of the cosmic rays in the free air from observations inside a closed vessel, account must be taken of the ionisation due to secondary radiation from the walls of the vessels. For a brass ionisation chamber with walls 2·5 mm. thick Hess has

shown that the secondary effect is only $0·09 I$.[6] It is therefore likely that the free-air value of q_h in temperate latitudes is not far from $1·90 I$.

The greater part of the residual ionisation arises from α rays and takes place close to the walls; q_0 is therefore the effect of the residual ionisation near the walls divided by the volume of the vessel.

9. Summary of Ionising Powers of Various Agencies. The table on page 14 gives a rough idea of the effects of the three main factors in the ionisation of the air at sea-level, both in closed vessels of steel, a few millimetres thick, and in the free atmosphere. The values given refer to measurements at about a metre from the surface of the earth; they are averages and, except in the case of the cosmic radiation, are subject to considerable fluctuations with local conditions.

It is noteworthy that the rate of production of ions over land areas is about four times as great as over the oceans, where the ionisation is practically all due to the action of the cosmic rays. Over land and close to the surface of the earth, these rays are much less important, but at greater heights the effects of the earth and the air-radiations fall off rapidly as we have seen, so that from a height of one or two kilometres upwards, the cosmic radiation is the chief agent in making the air a conductor.

10. The Rate of Disappearance of the Small Ions. The number of ions produced by the agencies we have discussed will not increase indefinitely as time goes on. Even in pure filtered air, free from nuclei, the small ions disappear by diffusion to the walls and by recombination with one another. Experiments have shown that when the diffusion loss can be neglected, the equation $dn/dt = q - \alpha n^2$ applies to such filtered air, n being as before the number of positive or negative small ions per c.c., q the number of pairs of ions produced per c.c. per second, and α the so-called recombination constant of small ions. In the course of time a state of equilibrium will be reached, in which the processes of creation and disappearance of ions balance one another; then $dn/dt = 0$

and the equilibrium equation $q = \alpha n^2$ holds. Applying this equation to the free air, taking $q = 7\cdot4\,I$ from the preceding section and $\alpha = 1\cdot6 \times 10^{-6}$ cm.^{3}/sec. from laboratory experiments, we find $n = 2145$. This is very much larger than the number of small ions actually observed, which is round about 600 per c.c. The discrepancy is due to neglect of the important part played by the Aitken or condensation nuclei in capturing the small ions.

TABLE I

IONISING POWERS AT 1 m. ABOVE THE EARTH'S SURFACE, SEA-LEVEL

Ionising Agent.	Ionising Power in Ion-pairs per c.c. per sec. in Air at N.T.P.			
	Unscreened Closed Vessel.		Free Air.	
	Over Land.	Over Sea.	Over Land.	Over Sea.
Radioactive matter in the earth . .	3·5	0	3·5	0
Radioactive matter in the air . . .	0·2	0	2·0	0
Cosmic Radiation .	2·0	2·0	1·9	1·9
Total ionising power for the free air .			7·4	1·9

This question has been examined over many years, principally by Nolan and his co-workers in Dublin, and it is found to be extremely complex.[3] Separate considerations must be given to the capture of small ions by charged nuclei of opposite sign, which leads to neutralisation, and by uncharged nuclei, which leads to the formation of large ions.

For the small ions, making no distinction of sign, we write :
$$dn/dt = q - \alpha n^2 - \eta_0 n N_0 - \eta n N \qquad . \quad (1.6)$$
where N_0 and N are the numbers of uncharged and of oppositely charged nuclei per c.c. and η_0 and η are the coefficients for the two types of capture. For the large ions :
$$dN/dt = \eta_0 n N_0 - \eta n N \qquad . \quad . \quad (1.7)$$
disregarding the small loss by recombination of large ions with one another.

For equilibrium both dn/dt and dN/dt are zero, hence $\eta_0 N_0 = \eta N$ and
$$q - \alpha n^2 - 2\eta n N = 0 \qquad . \quad . \quad (1.8)$$
From experiment it is known that η is of the same order of magnitude as α and since $N \gg n$ as a rule, eqn. (1.8) reduces to
$$q = 2\eta n N = 2\eta_0 n N_0 \qquad . \quad . \quad (1.9)$$
If the total number of nuclei, charged and uncharged, is Z, so that $Z = N_0 + 2N$, eqn. (1.9) reduces to
$$q = \left\{ \frac{2\eta_0 \eta}{2\eta_0 + \eta} \right\} Z n \qquad . \quad . \quad (1.10)$$
This is of the same form as an equation given originally by von Schweidler, $q = \beta n$, though β is constant only if the concentration of nuclei, Z, is constant.[5] It states a linear law of recombination, as opposed to the quadratic law which holds for filtered air. If the action of the ionising agents could be removed, the small ions would disappear at a rate $dn/dt = \beta n$, and $n = n_0 e^{-\beta t}$ would be the number present at any subsequent time. The quantity $\theta = 1/\beta$ can, by analogy with the equation for radioactive decay, be termed the *average life* of a small ion. Eqn. (1.10) then becomes
$$q\theta = n \qquad . \quad . \quad . \quad (1.11)$$
in which θ should vary inversely as the concentration of nuclei in the air. Though this simple relation is only an approximation, it is useful for general purposes.

The average life θ is found over land areas to lie between 10 and 60 sec. according to the purity of the air, the higher values being associated with observations made in the country. Over the oceans it may be as much as 230 sec., according to the measurements of Hess [7] and of Mathias [8] on the island of Heligoland. The table below, taken from Hess, illustrates the great difference in the equilibrium between small and large ions brought about when the wind direction changes; the first set of observations was made when the wind blew from the nuclei-laden European mainland, the second when it blew from the polar regions :

TABLE II

OBSERVATIONS OF THE STATE OF ATMOSPHERIC IONISATION (HESS, HELIGOLAND)

Wind.	No. of Small Ions per c.c.(n).		No. of Large Ions per c.c.(N).		Total No. of Nuclei.	Average Life of Small Ion, secs.	
	n_+	n_-	N_+	N_-	$Z = N_0 + N_+ + N_-$	θ_+	θ_-
S.W. (Land)	220	267	1480	1510	6300	22	22
N.N.W. (Arctic)	794	843	200	200	1100	115	204

These authors find that for values of θ lying between 36 and 100 sec., Schweidler's linear eqn. (1.11) is valid; in purer air, where θ is greater than 100 sec., the simple equation no longer holds, for the quadratic recombination term αn^2 then becomes significant.

Measurements with city air have, however, shown that the sizes of the nuclei vary over a wide range and that the coefficients η_0 and η in eqns. (1.6) and (1.7) show a corresponding wide range of variation. [3]

When, as in a city, the air is continually being provided with fresh nuclei, they do not usually reach a final state of equilibrium with the small ions. This would require a time of the order of 15 min. and stable conditions such

as are found only in mist and fog. A further complicating factor is the discovery that the meteorological conditions which cause large variations in the nature and the number of the nuclei also alter the value of q by altering the radio-active content of the air.

11. Methods of Determining the Average Life of a Small Ion. The average life, θ, of the small ions can be determined for a sample of air enclosed in an ionisation chamber by two methods suggested by Schweidler. [5] The first involves the separate measurement of the ionising power, q, of the ionising radiations, and of the number of small ions per c.c., n, and substitution in the equation $q\theta = n$. q is obtained from the saturation current, i, using the relation $i = qev$, where v is the volume of the vessel. The corrections to be applied have already been discussed (§ 8). The quantity n is found by allowing the air to stand undisturbed for a sufficient time for equilibrium between small ions and nuclei to be attained and then applying a strong field for a few seconds to sweep the small ions on to the collecting electrode. The second method involves the measurement of the currents flowing in an ionisation vessel for various values of the applied potential difference. [9]

12. The Ionisation-balance in the Lower Atmosphere. It is of considerable importance to decide to what extent the various factors in the creation and destruction of ions account for the state of ionisation actually observed in the air over land and sea. Any failure in this respect would suggest that new processes must be sought.

Although the variable nature of the process of capture by heavy nuclei prevents an accurate comparison of the rate of production of small ions with their rate of disappearance, the available evidence suggests that the chief factors in the ionisation-balance have been discovered. Over land areas where nuclei are usually plentiful, the Schweidler equilibrium equation, $q\theta = n$, applies approximately. An average value of 60 sec. for θ, the mean life of a small ion, and of 650 per c.c. for the small ion concentration, requires that the rate of creation, q,

should be 650/60 or 10·8 ions/c.c./sec. in reasonable agreement with direct measurements of q, shown in Table I. No direct measurements of θ have yet been made over the oceans. The concentration of nuclei in mid-ocean has been found to be from 400 to 800 per c.c. and the mean life-time of a small ion should be about 300 sec.[10] Since n was found on the *Carnegie* to amount to some 550 small ions/c.c., the ionising power at sea should amount to 550/300 or 1·83 I, in agreement with the value of q in the last column of Table I.

The problem of the ionisation of the lower atmosphere cannot be said to be completely solved until much more has been done to settle the manner in which the size and numbers of the nuclei affect their interaction with small ions. The general nature of the factors in the ionisation-balance has, however, been established, and an explanation can be given of the paradox that the conductivity of the air undergoes no large change as one passes from mid-ocean to land, despite a four-fold increase in the rate at which ions are formed. The increased birth-rate of the small ions is counterbalanced by an increased death-rate, owing to the greater concentration of condensation-nuclei over the land areas.

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CHAPTER II

ELECTRIC FIELDS AND ELECTRIC CURRENTS IN THE ATMOSPHERE

13. The Fine-weather Field. It was first found in the year 1752 by Lemonnier that the air above the earth is the seat of a persistent electric field during fine weather. The direction of this field shows that the earth carries a negative charge and the upper layers of the air a positive one. The state of affairs close to a flat portion of the surface of the earth can be expressed in three ways : (a) the earth carries a charge of surface density σ per unit area ; (b) there is a vertical field of strength $F = 4\pi\sigma$, just above the ground ; or (c) between two horizontal planes close to the ground there is a difference of potential

$$dV = V_{h+dh} - V_h = - Fdh = - 4\pi\sigma dh,$$

where $h + dh$ and h are the heights of the planes.

The quantity dV/dh is called the potential gradient and is positive, since σ is found to be negative. For the same reason, the field F is directed downwards, and this downward direction is by convention adopted as the positive direction for electric fields in the atmosphere. In these definitions it is assumed that the portion of the ground considered is flat and far removed from projections, such as trees and buildings, which would disturb the distribution of charge and concentrate the lines of force at certain points.

The average value of the fine-weather potential gradient is about 100 v./m. ; the corresponding value for the average charge density is $2·7 \times 10^{-4}$ electrostatic units/sq. cm. or 0·0009 coulomb/sq. km. The total fine-weather charge on the earth is of the order of 500,000 coulombs. Measurements of the fine-weather field are made by determining either σ or dV/dh ; they are always expressed in terms of the latter, in volts per metre.

14. Direct Measurement of the Surface Density of the Earth's Charge. The charge on such a limited portion of the ground as can be isolated for the determination of σ is so small that a sensitive electrometer must be used to measure it. Fig. 3 illustrates the "Universal
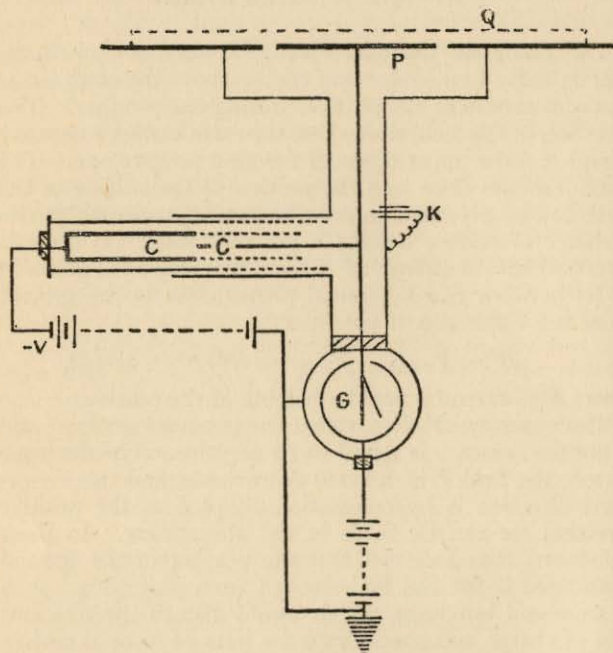

FIG. 3.—Universal portable electrometer (Wilson).

portable electrometer " devised by C. T. R. Wilson for this purpose. A flat circular plate P, called the test-plate, is surrounded by a guard-ring and mounted flush with the surface of the earth but insulated from it. This plate is joined to a gold-leaf electroscope G, whose leaf moves inside a positively charged case and is observed through a telemicroscope. C is a variable cylindrical condenser,

called a compensator, with its inner plate joined to the gold-leaf system, and its outer plate maintained at a constant potential, $-V$.

To make a measurement, the instrument is initially shielded from the earth's field by placing over it the earthed metal cover Q, indicated by the dotted lines. The reading of the leaf corresponding to earth or zero potential is found by momentarily earthing the plate P and the gold-leaf system with the key K. This earthing key is then withdrawn, the cover removed and the plate exposed to the field, with the compensator capacity zero. A negative charge thus appears by induction upon P, an equal positive charge is set free upon the central system, and the leaf moves inwards. It can be restored to its zero position if a negative charge is induced on the central system by increasing the capacity of the compensator to a value C'. The charge on the test-plate is then $-C'V$ and the surface density, σ, is $-C'V/A$, where A is the area of the test-plate.

In its original form the apparatus is small and portable, and the two batteries shown in the figure are replaced by small Leyden jars of silvered quartz, which keep their charge for a considerable time. For absolute measurements, with the plate mounted at ground level, a pit must be provided to house the observer ; very often relative measurements are made with the instrument mounted on a tripod, and the factor necessary to reduce the observations to standard conditions determined from simultaneous observations in a pit or with a potential gradient method. Since the lines of force are concentrated upon any projection above the earth, the factor is greater than unity ; in the case of a tripod 1 m. high it amounts to 3 or 4. A useful method of calibrating the instrument is to place it in artificial electric fields of known strength. Two large flat metal sheets, one flush with the test-plate and the other above it, are charged to a measured difference of potential by means of a battery of small cells.[1]

15. Measurement of the Potential Gradient. The potential gradient dV/dh may be found by measuring the

difference of potential between two insulated conductors at different heights above the ground. Various forms of electrometer are employed, and one conductor may be the earth itself, while the other is a stretched horizontal wire about a metre above ground-level. Sometimes two wires at different heights are used. It is of course essential that the supports of the wires and the presence of the observer and his instruments should not materially alter the field to be measured. In the arrangement in Fig. 4 (*a*), the wire is hung between amber or sulphur insulators at the ends of two vertical rods; its length should exceed five times its height above ground.
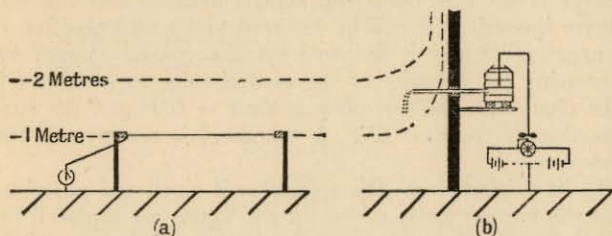
FIG. 4.—Measurement of the potential gradient.
(*a*) Absolute determination in open field.
(*b*) Continuous recording with the Kelvin water-dropper.

The wire when first set up is at a different potential to that of the air in its neighbourhood, and a current will flow between them which tends to remove this potential difference. The ordinary conductivity of the air is, however, too small for this process of equalisation to take place very rapidly and the electrometer reading will always lag behind a fluctuating field. It is necessary for this reason to place at the centre of the wire some active ionising agent which speeds up the process by increasing the local conductivity of the air. This may take the form of a glowing fuse of filter-paper, impregnated with a 5 per cent. solution of lead nitrate, or of a disc or spiral of metal coated with a deposit of polonium or radio-thorium giving off α rays. Such "collectors", as they are called, have

the advantage of causing the wire and the electrometer to acquire the potential of the air at the centre of the stretched length, where the electric field is least disturbed by supports and observer. The time taken for equalisation of potential depends upon the type and activity of the collector and varies from 1 to 60 sec. There is always some uncertainty as to the exact position at which the ionising agent exerts its effect, especially if a wind is blowing, but this can be avoided by making observations at different heights.

16. Continuous Recording of the Potential Gradient. An adaptation of the arrangement just described is employed to obtain continuous records of the potential gradient, the electrometer being housed in a building and the collector carried on an insulated rod projecting from the wall. The earth's field is very much distorted by the presence of the building, and a reduction factor for the installation must occasionally be determined by making simultaneous measurements in an open field. Fig. 4 (*b*) shows the nature of the distortion of the equipotential surfaces due to a building.

The Kelvin water-dropper is still sometimes used at recording stations as a potential equaliser. A jet of water from an insulated cistern within the building escapes from the end of a pipe passing through and insulated from the wall, and breaks up into fine drops at the point where the potential is required (Fig. 4). On starting the jet, the potential of the insulated system will in general be somewhat lower than that of the air at the end of the pipe; this end will carry an induced negative charge and the cistern a positive one. Each drop carries a negative charge away and leaves the system slightly higher in potential than before. Ultimately the negative charge at the end of the pipe disappears and the system is at the potential of the air at the point at which the drops break away. In practice this takes a time of the order of 30 sec.

Radioactive collectors are also employed, but whatever the type of collector, the insulated system is usually joined to the needle of a quadrant electrometer of low sensitivity, and opposite pairs of quadrants connected to a battery

whose centre is earthed. The record can be made intermittently, by means of a clockwork arrangement which depresses an inked pointer attached to the needle (Benndorf) or continuously, by receiving the spot of light from a small mirror on a moving strip of bromide paper.

17. The Electric Fluxmeter or Field Mill. If the cover Q in Fig. 3 were rapidly and periodically swung away from and over the plate P, the bound charge on P would alternately appear and disappear. If P is earthed through a high resistance R, an alternating potential difference of amplitude proportional to the field will appear across R. This can be amplified and if necessary rectified so that an output meter will show the magnitude of the field, an auxiliary device indicating its sign. Calibration can be carried out by comparison with a stretched wire in the open (§ 15).

Developments of this principle which was first employed by Russelvedt in Norway, have been used to measure electric fields during aircraft flights and under thunderclouds. In a recent form, capable of recording rapidly-varying fields, P is a ring of separate metal studs and Q is a rotating disc with a corresponding set of holes.[2]

18. Variation of the Potential Gradient with Height above the Ground : Space-charge. A good many observations of the potential gradient have been made with balloons fitted with collectors of the radioactive or the glowing fuse pattern ; they all indicate a rapid decrease in the field with increasing height above the ground (Table III, § 25). Though no great accuracy can be claimed for these measurements—for they are subject to errors arising from the distortion of the field by free and induced charges on the balloon and are but momentary observations of a quantity subject to considerable variation—it is well established that at a height of 18 km. the field has fallen to less than 1/100 of its value at ground level and is still diminishing slowly.

This diminution in the potential gradient with height indicates that a free positive charge, practically equal to the negative charge on the surface of the earth, resides

within the lower 18 km. of the atmosphere. Integration of the values of dV/dh obtained in balloon ascents gives a total potential difference between the earth and the air at a height of 18 km. of about 4×10^5 v. ; at this height, and above it, the conductivity of the air is so great that this figure may be taken to represent the potential of the highly conducting "Kennelly-Heaviside layer" some 80 km. above the ground.

The density, ρ, of the positive "space-charge" referred to can be determined from Poisson's equation, which, if the lines of force are vertical, takes the form

$$\frac{d}{dh}(dV/dh) = -4\pi\rho.$$

Several investigators have examined the nature of the space-charge quite close to the ground, using collectors at different heights. The results obtained are conflicting as to the magnitude, and even the sign, of ρ in this region ; evidently local conditions are of main importance.

19. Variations of the Potential Gradient with Locality, Time and Season. Before any discussion of the more important variations of the potential gradient, mention should be made of the fact that incessant fluctuations are observed over short periods of time ; these are due to local changes in the space-charge and the conductivity of the air in the vicinity of the collector.

For many years continuous records of the potential gradient have been taken at certain stations in the Northern Hemisphere. Less extensive information is available for the Southern Hemisphere, but observations have been made over considerable periods in the Arctic and Antarctic regions. Some very important measurements over the oceans have been carried out by the survey ship *Carnegie* of the Carnegie Institute of Washington. It appears that the sign of the potential gradient in fine weather is positive all over the earth's surface, but that over land areas its value varies considerably with local conditions. The mean of dV/dh at Kew is 317 v./m., while that at Davos in Switzerland is but 64 v./m. Over the oceans a value of

126 v./m., which varies little if at all with geographical position, has been found. The average value for the whole earth is not far from 120 v./m.

It would seem that the potential gradient observed on land is not a quantity of much fundamental significance, its value depending primarily upon local conditions of atmospheric conductivity. This is evident from the fact that a simple periodic variation according to *local* time is the basis of all fine-weather land records. The minimum
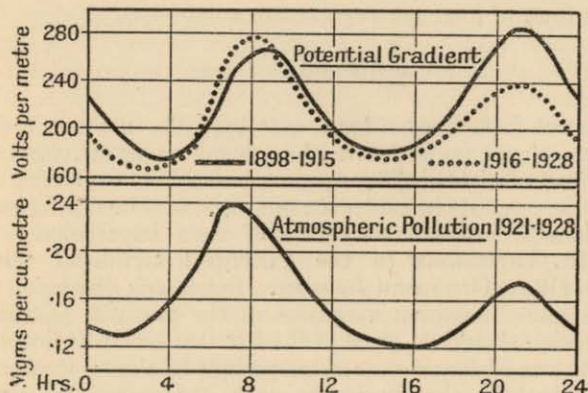

Fig. 5.—Relation between potential gradient and atmospheric pollution (Whipple).

values are obtained in the early morning at about 4 a.m. and the maxima in the evening between 6 and 8 p.m. In many places a second maximum at 8 a.m. and a second minimum at midday are found. The amplitude of these daily variations at land stations sometimes reaches 50 per cent. of the mean value for the day.

The local time variation has been shown by Whipple [3] to be closely correlated with the amount and distribution of the smoke pollution of the atmosphere near large cities, where these continuous records have generally been made. Fig. 5 shows the variation at Kew of the potential gradient during the summer months for the two periods 1898-1915

and 1916-28, together with the variation in the amount of pollution (determined by aspirating the air through filter-paper) for the period 1921-28. The figure shows not only the close relation which exists between potential gradient and pollution but also the striking effect of the introduction of " summer time " into Great Britain in the year 1916, after which the morning minimum and maximum moved back by approximately one real hour. The atmospheric pollution curve has maxima and minima which have been explained by Simpson as due to the combined effect of variations in the amount of smoke produced in the city and variations in the stability of the atmosphere, that is in the mixing action of surface winds and general turbulence of the lower air.

As regards the annual variation of the potential gradient, land stations in both hemispheres show a maximum in the local winter and a minimum in the local summer. The only significant exception is the Antarctic region, where a reversal of phase occurs, giving rise to a maximum in the local summer and a minimum in the winter.

In view of the local origin of the variations of the gradient found at land stations near towns, observations over the oceans, where no pronounced local effects are to be expected, are of very great importance. Mauchly [4] analysed the measurements of the *Carnegie* cruises and found that there is a well-marked diurnal variation at sea, with maxima and minima occurring at the same moment in all parts of the globe. Thus the potential gradient in all oceans is found to be about $15\frac{1}{2}$ per cent. below the mean at 5 hours, G.M.T., and about 20 per cent. above it at 19 hours, G.M.T. It is remarkable that the same 24-hour wave, progressing according to universal solar time, had been observed by Simpson in the Antarctic and Lapland. Fig. 6 (a) (after Whipple) illustrates these results, which were confirmed by the last cruise of the *Carnegie*. In each case shown in the figure, the mean potential gradient has been reduced to 100 v./m. No appreciable annual variation has so far been observed over the sea.

Mauchly's discovery of a variation of potential gradient according to universal time is generally accepted as being of fundamental significance. The variation must clearly be ascribed to an alteration in the difference of potential between the upper conducting layers of the atmosphere and the earth; it affords an important means of testing any theory of the origin and maintenance of this potential difference.
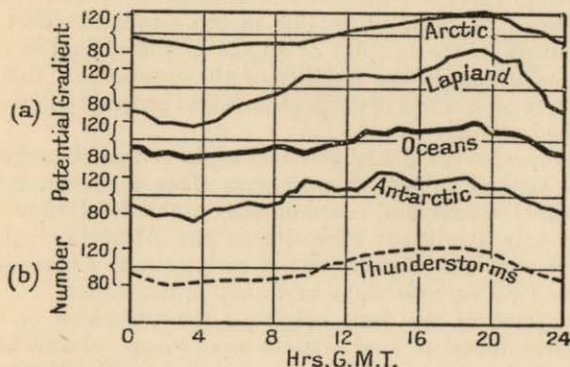

FIG. 6.

(a) Variation of potential gradient with universal time.
(b) Variation of thunderstorm activity over the earth's surface (Whipple).

20. The Fine-weather Current: Conduction and Convection Currents.

In the ordinary fine-weather field positive ions are driven towards the earth and negative ions away from it; the motion of these ions constitutes a downwardly directed conduction current. The ionic conduction current passing through 1 sq. cm. of a horizontal plane can be written as $I_0 = F(\lambda_+ + \lambda_-)$, where F is the strength of the electric field and λ_+ and λ_- are the polar conductivities of the air at the point in question (§ 3).

There is, however, a fine-weather current of a different nature which plays a part in the transfer of electricity by the atmosphere. If the air should contain at any point

an excess of ions of one sign (a space-charge), movement due to wind or ordinary turbulence will give rise to a mechanical transference of electric charge. Thus if v is the upward vertical component of the velocity of the air and ρ the space-charge per c.c., the upward convection current due to this cause will be $v\rho$ per sq. cm. The real current-density in fine weather is thus the resultant of the downward conduction and the upward convection currents, and is given by $I = I_0 - v\rho$.

The mean value of the total fine-weather current over the whole globe is not far from 2×10^{-16} amp./sq. cm. or 2 microamp./sq. km., so that the total current flowing in this way between the upper atmosphere and the whole earth is about 1000 amp. It would seem, from measurements which have been made in various parts of the world, that this current varies much less with changes in geographical position, in time of day and season of the year, than does the potential gradient. Although in the past it has not received a great deal of attention, it would appear to be a more useful quantity than the potential gradient and less affected by purely local conditions.

21. Direct Determination of the Conduction Current at the Surface of the Earth.

Direct measurements of the magnitude of the conduction current which flows into the earth's surface have been made with the Wilson universal electrometer, with the test-plate mounted flush with the surface of the ground, as described in § 14. The curve of Fig. 7 illustrates the procedure followed in the measurements. At time t_1 the test-plate is exposed to the earth's field by removing the cover, and the capacity of the compensating condenser is adjusted to the value C necessary to bring the gold leaf back to the zero position. The charge on the plate is then $-CV$ and the field strength $F = 4\pi \dfrac{CV}{A}$, where, as before, A is the area of the plate and $-V$ the potential of the outside of the compensator. The test-plate is now left exposed until time t_2 and kept continuously at earth potential by increasing the compensator capacity to balance the charge received

from the positive ions driven from the air on to the plate. Owing to variations in the strength of the field, and so in the induced charge on the plate, this increase is not a regular one. At time t_2 the plate is again shielded with the cover and the compensator capacity is reduced from C_1 to C_2 to maintain the shielded system at earth potential. At this moment the field strength is $4\pi(C_1 - C_2)V/A$. Since at the beginning of the measurement the compensator capacity was zero, it is evident that during the interval, $t_2 - t_1$, the plate received a positive charge C_2V
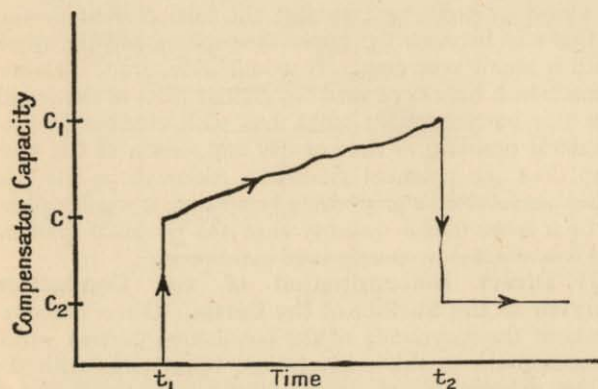
FIG. 7.—Measurement of conduction current with the Wilson universal electrometer.

and the average value of the conduction current per square centimetre was $C_2V/A(t_2 - t_1)$. For routine observations the instrument is mounted on a tripod above the ground, and the readings are reduced to standard conditions by occasional determinations of the reduction factor, from observations in a pit.

22. Indirect Determination of the Conduction Current at a Point in the Air. The conduction current at a point in the air can be determined also by separate measurements of the field strength, F, and the polar conductivities, λ_+ and λ_-. The methods used to determine

F, or the potential gradient, have already been described in this chapter; the conductivities are usually found by the Gerdien method sketched in § 2. In another form of the Gerdien method, due to Schering, the cylindrical condenser is replaced by a large earthed wire cage, along the axis of which is a charged insulated wire connected to an electrometer. The rate at which the charge is dissipated by ions of opposite sign is determined in exactly the same manner as in the Gerdien method, and the same equations apply if certain conditions are fulfilled. The cage must be large enough, and the charge on the wire small enough, to ensure that no considerable alteration in the number of ions is caused by the flow of current. The arrangement must be set up in the open air to ensure that ordinary atmospheric circulation prevents the development of the electrode space-charges to be discussed in the next section. And finally, the whole cage must be shielded from the earth's field by a rough roof or by trees, otherwise it will carry an induced negative charge on the outside and rob the entering air of negative ions.

23. The Comparison of the Direct and Indirect Methods : Electrode Space-charge. The direct method due to Wilson measures the conduction current entering the earth; the indirect method measures the current at a certain height above the earth. In the latter case the current is carried by two oppositely moving streams of ions of opposite sign; in the former it consists of a single stream, for while positive ions enter the earth, negative ions do not pass from it into the atmosphere. The two methods, therefore, do not necessarily measure the same quantity. One determines the true conduction current at a height, and the other its positive component at ground-level.

Consider a vertical cylinder of perfectly still air whose height is h and cross-section 1 sq. cm., with its base on the ground (Fig. 8). Let F' and F be the electric field strengths at top and bottom of this cylinder, and λ'_+, λ'_- and λ_+, λ_- the polar conductivities. The conduction current through the top is $F'(\lambda'_+ + \lambda'_-)$; in each second

a positive charge $F'\lambda'_+$ enters, and a negative charge $F'\lambda'_-$ leaves. Through the bottom there is only a downward flow of positive ions, carrying away a positive charge $F\lambda_+$ per sec. The cylinder as a whole thus loses a positive charge $F\lambda_+ - F'\lambda'_+$ per sec., and a negative charge $F'\lambda'_-$ per sec. This may be expressed as a development of a positive space-charge at a rate

$$F'\lambda'_- - (F\lambda_+ - F'\lambda'_+) = F'(\lambda'_+ + \lambda'_-) - F\lambda_+$$

per sec. The growth of this "electrode" space-charge should automatically decrease the field F' and should cease only when F' has such a value that as many negative as positive ions leave the cylinder in each second, i.e. when

$$F'(\lambda'_+ + \lambda'_-) = F\lambda_+ . \quad (2.1)$$

The two sides of this last equation are respectively the conduction currents at height h and at ground level. Thus, for perfectly still air, the two methods we have described should give the same numerical values, though they measure different quantities.

It was thought for some time that λ_+ must necessarily be considerably smaller than $(\lambda'_+ + \lambda'_-)$ and hence that F must be about double F'. Scrase[5] has, however, found practically no difference between F and F' and very little sign of electrode space-charge. A suggestion by Whipple that the space-charge is removed by atmospheric turbulence as rapidly as it is formed is unable to explain these results.
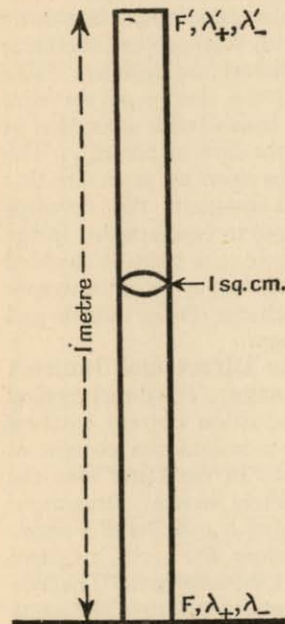
FIG. 8.—Electrode space-charge.

The difficulty appears to have been removed by the discovery by Hogg[6] that, owing to the local effect of α and β rays from radioactive matter in the soil, λ_+ is much larger than was thought and approximately equal to $(\lambda'_+ + \lambda'_-)$. Since $F \cong F'$, eqn. 2.1 is satisfied and the problem of electrode space-charge does not arise.

Values of the air-earth current in fine-weather obtained by the direct method in various parts of the world range from 1·0 to 4·0 × 10^{-16} amp./cm.^{2} Those by the indirect method vary from 0·4 to 3·2 × 10^{-16} amp./cm.^{2} Though as indicated above and found in practice, the two methods should give the same result, the variability of the ionising agents producing λ_+ and of convection currents suggests that the direct method is the more reliable of the two.

24. Variations in the Fine-weather Current. Both direct and indirect methods of studying the air-earth current indicate that in fairly fine weather it is less subject to variations than is the potential gradient. Simultaneous measurements of the field and of the total conductivity of the air, $\lambda = \lambda_+ + \lambda_-$, show that these quantities vary to some extent in inverse ratio, with the result that their product, the conduction current, is less variable than either. Factors such as an increase in the number of condensation nuclei or the presence of fog particles, which decrease λ by raising the proportion of the slow ions present, cause a corresponding increase in the potential gradient. The simple form of the daily variation in the earth's field has its counterpart in an inverse variation of the conductivity, with a maximum in the early morning between 3 and 4 a.m. The extreme effect of atmospheric pollution near large cities, which increases the potential gradient and diminishes the conductivity, has already been discussed (§ 19). In general however the reciprocal relation between λ and F is not exact enough to keep the conduction current invariable. At land stations in Europe the current is found to have an annual variation which coincides in phase with that of the potential gradient but has a smaller amplitude. At other stations its daily variation follows the changes in the conductivity.

25. The Conduction Current at Great Heights.
It has been found from simultaneous determinations of
potential gradient and atmospheric conductivity during
balloon ascents that the rapid decrease in F with height
above the ground is accompanied by an increase in λ,
of such magnitude as to keep the conduction current
approximately constant. Table III shows the results of an
ascent by Wigand,[7] from which it will be seen that when
the field had fallen to $1/15$ of its value at the ground, the
conductivity had risen eleven-fold.

TABLE III

CONDUCTION CURRENT DURING BALLOON ASCENT (WIGAND)

Height (Metres).	F (volts/metre).	λ (e.s.u.).	I (amps./sq. cm.).
0	136	$1 \cdot 1 \; 10^{-4}$	$1 \cdot 7 \; 10^{-16}$
2500	27	$4 \cdot 8$,,	$1 \cdot 4$,,
4400	18	$8 \cdot 2$,,	$1 \cdot 6$,,
6500	$8 \cdot 8$	$12 \cdot 6$,,	$1 \cdot 2$,,

These results have been confirmed and extended by
conductivity measurements made during the flight of the
United States army balloon Explorer II in 1935.[8] They
showed that up to a height of 18 km., the ionisation of the
air increases by roughly 500 ion-pairs per km. and that
the conductivity at 18 km. was about 100 times the usual
value at the ground.

26. The Significance of the Air-earth Current. The
general constancy of the air-earth current with height above
the ground and in different parts of the world suggests
that, at any rate to a first approximation, it has its origin
in a constant potential difference between the conducting
layers of the upper atmosphere and the earth, and that its
actual value depends upon the resistance of the air between
the two regions. The conductivity of the upper portion
of this air-path is entirely due to the cosmic radiation *

* This refers to the regions from 15 to 30 kilometres above
the ground. The E layer 80 kilometres up has a solar origin.

and is unlikely to alter appreciably with time of day,
locality or season. But below a height of about 2 km.
over land areas, an additional ionising influence is exercised
by radioactive matter in the air and in the earth, and this,
as we have seen (§ 5), varies very much. Moreover, the
conductivity of the air in this region is dependent upon
the relative concentrations of the large and the small
ions, and these alter with the number of nuclei present.
The current-bearing column of air above the point of
observation thus consists of an upper constant resistance
and a lower variable one in series with each other. The
extent to which the current carried by the air-column
alters with local conditions will depend on the ratio
which the lower variable resistance bears to the total
resistance.

Both the results of Wigand in Table III and those from
the Explorer II flight [8] indicate that the total resistance of
a column of air of cross-sectional area 1 sq. cm. is about
$1 \cdot 0 \times 10^{21}$ ohms. The conductivity between 14 and 18
km. is so high that this upper region contributes only a
small fraction to the resistance and higher regions would
be still less important. If the variable part of the column
be the first 100 m. above the earth, where the specific
resistance is usually 8×10^{15} ohm-cm., this portion will
contribute 8×10^{19} ohms to the total. We see that even
if the conductivity of the first 100 m. were halved, the
effect upon the air-earth current would amount to but
4 per cent. The potential gradient near the ground, on
the other hand, would be doubled, for from this point of
view it represents the fall of potential per unit length of
the lowest part of the current-bearing column. It is
evident that potential gradient measurements over land,
and in particular near cities, cannot be expected to yield
information which bears directly upon the question of
the total potential difference between upper atmosphere
and earth. In the case of observations at sea and in polar
regions the position is different ; practically all the ionisa-
tion throughout the whole column is then due to the
cosmic radiation, and the only local effect which can

influence the fine-weather gradient is a change in the number of nuclei available for the capture of small ions.

27. The Electric Field during Disturbed Weather. The normal downwardly directed field, due to a positively charged upper atmosphere and a negatively charged earth, is frequently disturbed when the weather ceases to be fine. During fog, as explained in § 24, it is very much increased, and may reach ten times its normal value; during dust-storms, of the type common in semi-arid regions and deserts, powerful reversed or negative fields are usual and reach 10,000 v./m. The effects found during cloudy weather and rain are variable, and range from gradients of the order of a few hundred volts per metre in a fine drizzle to as much as 50,000 v./m. under thunderclouds.

Generally speaking, negative fields predominate when light or steady rain is falling, though occasional positive excursions are observed. In the case of heavy rainstorms and thunderclouds, the sign of the field depends upon the portion of the cloud passing over the point of observation, but here, too, there is evidence that negative gradients are the more frequent.

The field during thunderstorms fluctuates a great deal if the cloud is active in producing lightning. Its value at the ground does not as a rule exceed 10,000 to 20,000 v./m. except for very brief intervals occupying fractions of a second. Equally high fields are observed during heavy rainstorms unaccompanied by lightning. During a snowstorm the field is usually positive and in a heavy storm may attain 10,000 v./m. The methods employed for the study of the large and rapidly varying fields of thunderclouds are described in Chapter III.

28. The Charge on Rain and Snow; the Precipitation Current. The apparatus required to examine the charge carried to the earth by precipitation—rain, hail, snow, and sleet—is comparatively simple, and consists essentially of an insulated metal vessel to catch the rain, etc., and an electrometer to measure the charge received. Great precautions must, however, be taken

to eliminate the disturbing effect of the negative charge given to the air as the drops splash on the container (Lenard effect). If some of this is lost a spurious positive charge will be indicated. For a similar reason, the entry of spray due to the splashing of rain falling outside the receiving vessel must be prevented. The receiver must be screened electrostatically from the electric field outside. The rate of rainfall, and sometimes the size of individual drops, is also observed. By amplifying the induction effect of a drop as it passes through an insulated cylinder the charges carried by individual drops may be measured. As the charges brought down vary considerably, it is necessary in all such work to carry out a long series of observations.

Since the year 1908, when the modern series of observations began, all observers have found a preponderance of positive charge on rain in general. The average ratio of the quantities of positive and negative electricity brought down by all kinds of precipitation varies according to the situation of the observer. Some of the values found for this ratio are: Potsdam (Schindelhauer), 1·4 : 1; Puy-en-Velay (Baldit), 2·4 : 1; Dublin (McClelland and Nolan), 4·8 : 1; Simla (Simpson), 2·4 : 1; Otago, New Zealand (Marwick), 3·2 : 1. As exceptions to the general rule, it may be noted that snow usually carries a considerable negative charge, and that the fine drops of a drizzle are also negative.

The subject of the charge on rain and its relation to the prevailing electric field has recently received special attention because of its connection with the electrification of thunderclouds. This work is discussed in § 45.

The charge per c.c. in the case of heavy thunderstorm rains reaches average values of 30 to 40 e.s.u., while single drops may carry 100 to 200 e.s.u. (Gschwend) and attain potentials as great as 300 v. The mean values reported by different observers for the net charge per c.c. on all forms of precipitation during a long series of observations range from + 0·029 e.s.u. at Potsdam (Schindelhauer) to + 0·176 e.s.u. at Simla (Simpson).

The transfer of electricity from atmosphere to earth in this manner constitutes a " precipitation " current which appears on the whole to be in the same direction as the fine-weather conduction current, since it carries more positive than negative charge to the earth. In the case of thunderstorms this current sometimes reaches values as high as 2×10^{-11} amp./sq. cm., and the total convection current carried by the rain in such a storm may amount to more than 0·1 amp. Before it is possible to frame a trustworthy estimate of the amount of the excess of positive over negative electricity conveyed to the whole earth in this way, much more information is required from tropical and equatorial regions.

29. Currents due to Point-discharge. It is well known that an electric glow or brush discharge takes place from a pointed conductor when placed in a sufficiently strong field. It consists of a stream of ions of the same sign as the charge upon the point. Discharges of this kind must, therefore, be expected from the ends of exposed conductors on the surface of the earth whenever the potential gradient and the height and sharpness of the ends are sufficiently great. Indeed the earliest investigators, Franklin, Dalibard and Lemonnier with high metal rods and kites detected electric fields in thundery and in fine weather, by means of this current. Under favourable conditions, afforded by mountain peaks or the masts of ships exposed to the intense fields of a thundercloud, the glow discharge may become very conspicuous (St. Elmo's fire).

It was first pointed out by C. T. R. Wilson [9] that point-discharge currents of much smaller intensity must play an important part in the interchange of electricity between the earth and the atmosphere. The fields prevailing under rainstorms and thunderclouds very frequently reach values sufficiently great for considerable currents to be discharged from an assemblage of exposed conductors, such as trees, bushes, housetops and even fields of grass. A conductor need not end in a sharp point nor project to a great height in order that it should begin to act as a discharger. For example, an earth-connected sphere of radius 1 cm. need

be raised only to a height of 3 m. in the not unusual field of 10,000 v./m. for the electric intensity at its surface to reach the critical value of 30,000 v./cm. at which brush discharge begins. A practical instance of the correctness of this suggestion is afforded by some experiments made by the author, in which a small tree, 4 m. high, was cut off at its base, mounted on insulators and connected to earth through a galvanometer. Exposure to the fields due to nearby thunderstorms yielded the average point-discharge currents shown below : [10]

TABLE IV

POINT-DISCHARGE CURRENTS FROM A SMALL TREE (SCHONLAND)

Field (volts/metre).	Current (microamperes).
— 3,500	0·07
— 5,500	0·20
— 11,000	1·00
— 16,000	4·00

The contribution of a single tree such as this must be multiplied by a very large factor to obtain the total current flowing between the earth and a thundercloud ; in the case quoted, the tree was typical of the exposed natural conductors for many miles around, and a rough surface integration over the area affected by the cloud indicated that the total point-discharge effect was of the order of 2 amp. in an upward direction.

Since the direction and magnitude of the large fields associated with disturbed weather vary considerably, it is important to know the relative amounts of positive and negative electricity discharged from an exposed point over a long period of time. In a series of observations by Wormell,[11] a sharp point carried on insulators at the top of a pole was connected by cable to one electrode of a voltameter filled with dilute sulphuric acid, the other

electrode being earthed. The voltameter, shown, with some minor but significant details omitted, in Fig. 9, was constructed of fine capillary tubing (diameter 0·8 mm.) and the gases evolved were separately collected, their volumes being determined from the lengths of the bubbles formed. If v_1 and v_2 are the volumes of the mixed gases collected at the earthed and the point-connected electrodes respectively, the quantities of positive and negative electricity, q_1 and q_2 which have been discharged from the point can be separately determined in the following way : If one unit of quantity liberates in electrolysis c c.c. of oxygen and $2c$ c.c. of hydrogen, we have

$$q_1 c + 2q_2 c = v_1 \quad \text{and} \quad 2q_1 c + q_2 c = v_2,$$

from which we may determine q_1 and q_2.

H₂SO₄ *dil.*

Mercury   Pt.Wire Gas
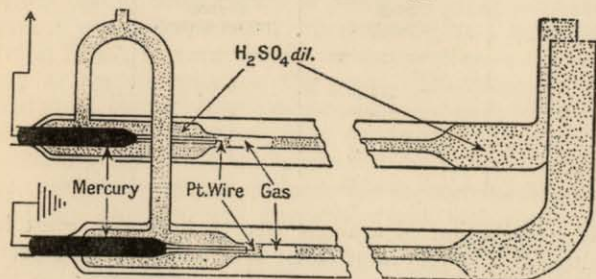
FIG. 9.—Water micro-voltameter (Wormell).

The quantities discharged in this way have been examined by Wormell at Cambridge over a period of three years. Some of his results are shown in the Table V.

Each year shows a considerable net loss of positive electricity from the discharger.

In a more detailed discussion it is noted that of 147 separate periods of disturbed weather accompanied by precipitation, 103 showed an excess of upward current (+ ve discharge), 34 an excess of downward current (− ve discharge), and in 10 the net discharge was zero. This is what would be expected from the predominance of negative potential gradients in disturbed weather.

Similar measurements have been made in other parts of the world (§ 30).

TABLE V

INTEGRATED EFFECTS OF POINT-DISCHARGE (WORMELL)

Year	1927.	1928.	1929.
Positive quantity discharged, q_1 (coulombs)	0·28	0·24	—
Negative quantity discharged, q_2 (coulombs)	0·14	0·11	—
Net positive discharge, $q_1 - q_2$ (coulombs) .	0·14	0·13	0·11

30. The Interchange of Electricity Between the Earth and the Atmosphere ; the Maintenance of the Earth's Charge. The negatively charged earth and the positively charged upper layers of the atmosphere form two plates of a spherical condenser with the lower air as the dielectric. Although the conductivity of the air between the plates is small, the applied potential difference is great enough to make the leakage of charge through it very considerable. The average value of the charge per unit area of the earth's surface is 9×10^{-14} coulombs/sq. cm., and that of the fine-weather conduction current is 2×10^{-16} amp./sq. cm. Left to itself, a condenser of this kind would be discharged by internal leakage in a time of the order of 10 min. But this is not all ; to the dissipation of the earth's charge by the fine-weather current we must add the effect of the charge conveyed by rain and snow, for it has been shown that on the whole the net charge conveyed to the earth by precipitation is positive in sign. Thus to the conduction current of 1000 amp. over the whole surface of the earth must be added a precipitation current in the same direction, estimated by Wigand [12]—though any such estimate can

only be provisional—at 400 amp., making a total of 1400 amp. tending to dissipate the 500,000 coulombs with which the earth is charged.

In spite of the continuous operation of these two factors, the earth's charge remains practically constant, so it is clear that there must exist some reverse or compensatory process which neutralises their effect. Some agency must be continuously at work replenishing the charges on the earth and in the upper air. The discovery of the mechanism responsible for this replenishment has been one of the chief problems of atmospheric electricity, and many proposals have been made to this end.

It was first suggested by C. T. R. Wilson [13] that it is to regions of disturbed weather that we must look for the mechanism of replenishment, that under rain and thunderclouds, where the potential gradient is more often reversed than in the normal direction, there are two processes at work which convey considerable quantities of negative charge to the earth. The first of these is the action of point-discharge from conductors projecting from the ground. This has already been discussed, and it has been shown that experimental tests indicate that it is of great importance. The second process is the charge conveyed to the ground by lightning flashes from thunderclouds. Every second, as will be seen in Chapter III, some 100 lightning discharges occur over the whole surface of the earth, each involving the passage of a charge of the order of 30 coulombs. Only a fraction of these, perhaps one in four on the average, strike the ground. The effectiveness of this process of transference of charge will depend upon whether more charges of one sign than another are conveyed to the earth. If, for example, practically all the charges transferred were negative, the process would be equivalent to a continuous compensation current of about 500 amp. Though there is evidence from a number of regions (§ 33) that this is actually the case, the question is still under investigation.

An estimate of the annual electrical " balance-sheet " of a square kilometre of ground at Cambridge in England

has been made by Wormell,[11] on the basis of his point-discharge measurements and observations on lightning discharges. The air-earth conduction current loss and the effect of rain and snow were estimated from average values of these quantities. The results were :

	Coulombs/sq. km./annum.
By natural point-discharge, gained . . .	— 100
By lightning, gained	— 20
By atmospheric conduction, lost . . .	— 60
By precipitation, lost	— 20
Net gain of negative charge	— 40

Though such an estimate is only a very rough one, it appears to be quite possible that in this locality the four processes approximately balance one another, or even that the earth gains a negative charge.

These results have been supported and extended by similar measurements at Kew Observatory, at Durham, at Pretoria and in Nigeria.

Over the oceans, where thunderstorms are infrequent and elevated points are absent, the balance-sheet may be expected to show a net loss of negative charge, so that the position for the whole globe would be one of approximate equilibrium.

This suggestion of Wilson involved more than the interchange of electricity between the bases of cumulonimbus clouds, showers and thunderstorms, and the ground ; he regarded such clouds as equally active in supplying positive electricity to the upper air. The existence of this upward current has been established by recent high-altitude flights above thunderclouds. Gish and Wait,[14] by measuring both the electric field and the conductivity of the air, have shown that an average current of 0·8 amp. flows in the expected direction. The cloud thus acts as an electrical generator which removes positive electricity from the earth and supplies it to the conducting layers of the atmosphere above, in which it is rapidly distributed in

such a way as to maintain them at a constant potential of about 4×10^6 v., in spite of losses.

We should therefore expect the potential difference, and so the potential gradient, to show a maximum when the thundery regions of the earth are at their maximum activity. The diurnal variation of the gradient discovered by Mauchly (§ 19) should agree in phase with the diurnal variation of thunderstorm activity over the earth. It has been shown by Whipple [15] that this appears to be the case. Whipple's examination of the diurnal variation of the world's thunderstorms is represented by the bottom curve of Fig. 6 (b), where it will be seen that the parallelism with the variation of the potential gradient over the oceans and the polar caps is indeed very close.

Although the evidence discussed above makes it very probable that Wilson's replenishment theory is correct, it must be admitted that further quantitative measurements in different parts of the globe are needed to establish it.

A rather revolutionary extension of Wilson's ideas has been put forward by Frenkel,[16] who suggests that the generators chiefly responsible for the circulation of electricity between the earth and the upper air, in the reverse direction to the fine-weather current, are not thunderstorms but ordinary clouds. According to this theory all fog-like clouds carry a positive charge above and a negative charge below (§ 44) and thus create a conduction current which is opposite to that of fine weather. The fields due to these cloud dipoles (of which there are estimated to be a hundred thousand at any moment over the earth's surface) will, in Frenkel's view, induce and maintain the negative charges observed on the distant " serene sky " regions of the earth and positive charges in the air above them.

It has been shown, however, in § 24, that the enhanced fields below ordinary fog-like clouds are due to an alteration in the conductivity of the air and are in the normal positive direction. Such clouds do not give reversed electrical fields of the kind required by Frenkel's theory unless they are shower-clouds from which rain is falling (§ 45).

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CHAPTER III

THE ELECTRIFICATION OF THUNDERCLOUDS

31. The Number of Thunderstorms over the Globe. Although a thunderstorm is a comparatively rare event at any one station in temperate latitudes, the number of thunderstorms per day experienced by the earth as a whole is very large. Even in such a limited area as France there are comparatively few days in the year on which thunder is not reported from some part of the country. According to a survey made by Brookes,[1] the earth experiences 16,000,000 thunderstorms per annum or 44,000 per day. Taking one hour as the average duration of a single storm there will on the average be 1800 thunderstorms in progress in different parts of the world at any one moment. As has been shown in § 19, Fig. 6, world thunderstorm activity alters regularly with the sun, reaching a maximum when the afternoon storms of equatorial Africa are ending and those of equatorial America are beginning. Brookes has estimated that about 100 flashes of lightning occur every second or 360,000 every hour.

The most thundery region of the earth is the island of Java, where thunder is heard at any one place on 61 per cent. of the days in a year. This percentage, the *isoceraunic level* of the place concerned, is 41 in parts of Central Africa, 39 to 37 in Southern Mexico and Panama, 29 in Brazil and 26 in Madagascar. In North America the isoceraunic level rises from 3 per cent. for Southern Canada to 25 per cent. for Florida. On the Californian coast it is about 1 per cent.

Over most of Europe thunder is heard on the average on about 11 days in the year; near the Alps the figure rises to 18 days. The figure for Johannesburg in South Africa is 50-60 days in the year. In using these figures account must be taken of the fact that on about 1 in 7 of thunderstorm "days" there is more than one storm within audible distance.

Over the oceans thunderstorms are infrequent except off the south-east coasts of Brazil and South Africa and in the seas surrounding Madagascar and the East Indian islands.

32. Methods used to Measure the Electric Fields of Thunderclouds. (*a*) *The capillary electrometer method.* This was devised by C. T. R. Wilson[2] for use in his classic investigations of thunderstorms and has also been employed by Wormell[3] and by the writer.[4] The general principle is similar to that of the test-plate or induced charge method of studying the fine-weather field (§ 14), in which an exposed conductor is kept at zero potential by alteration in the capacity of a compensating condenser, when the charge induced upon the conductor by the field is equal to that given by the compensator to the system. In the measurement of rapidly changing thunderstorm fields, however, compensation is produced automatically and very quickly by means of an ingenious form of capillary electrometer. As shown in Fig. 10, this consists of a small bubble of dilute sulphuric acid enclosed between mercury threads in a narrow capillary tube. The threads lead to end-cups of mercury, and one of these, A, is connected to a test-plate T insulated from, but flush with, the surrounding ground, while the other is connected to the earth. Between the mercury and the glass there is everywhere a thin film of dilute acid, so that an electrical "double layer" surrounds the mercury and makes each half of the electrometer a condenser charged to the potential difference of the double layer, about 0·90 v., the mercury being positively charged with respect to the acid. If a positive charge is given to the left-hand side, it will increase the charge on the left-hand condenser and the left-hand thread will move forward a certain distance so as to absorb this charge, by increasing the area of the acid-mercury surface and the capacity of the left-hand condenser. A corresponding movement of the right-hand thread causes a reduction in area on this side and the passage to earth of a quantity of electricity exactly equal to that originally supplied to the electrometer. The distance moved, *y*,

is easily seen to be related to the charge passed through, q, by the equation

$$q = 2\pi r C V'(1 - r/R)y,$$

where r is the radius of the capillary and R that of the end-tube, and V' is the potential difference, and C the capacity per unit area, of the double layer. Thus $q = ky$, where k is an instrumental constant which can be determined by discharging a known quantity of electricity through the instrument. The arrangement, therefore, gives a linear relation between the movement of the acid bubble and the quantity discharged. In spite of the large
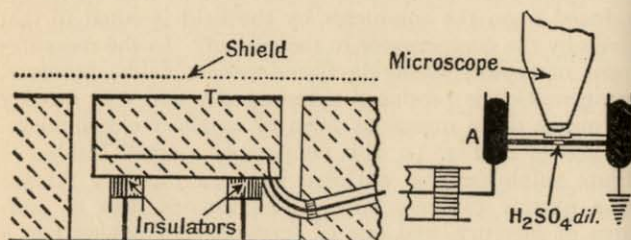
FIG. 10.—Measurements of the electric fields due to thunderstorms (C. T. R. Wilson).

value of C, about 30 microfarads/sq. cm., the electrometer is extremely sensitive. By using a microscope focused upon one end of the bubble, quantities of the order of 10^{-9} coulombs may be measured. A permanent record is best obtained by placing a slit and a moving photographic plate in the focal plane of the objective.

When connected to a test-plate the electrometer automatically maintains it at earth potential and records the quantities of electricity passing to and from the earth as a result of changes in the external field and of the induced charges on the plate. The type of record obtained from this arrangement during a thunderstorm is illustrated by the thick line of Fig. 11. An earthed cover is swung over the test-plate for a few seconds

at the beginning of each minute, causing the movements A'AA", B'BB", etc., as the induced charge flows away to earth and then returns. The points A, B, C, D, E would lie on a horizontal line if there were no actual transfer of charge between the plate and the atmosphere taking place, that is to say, no conduction current, point-discharge, or transport of charge by rain. The first 2 min. in the figure show the slight slope due to the conduction current only, the last two show the effect of a fall of positively charged
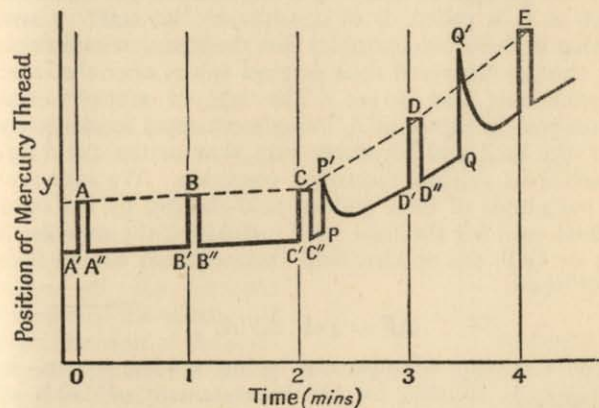
FIG. 11.—Capillary electrometer record during a thunderstorm.

rain. The charge, for example, conveyed to the plate by rain and conduction during the third minute is $k(y_D - y_C)$ and the mean value of the current per square centimetre $k(y_D - y_C)/60A$, where A is the area of the plate and k the constant of the electrometer. The dotted line joining the points A, B, C, D, E gives the reading of the instrument when the test-plate is shielded at any time during the record, and this line serves as the zero from which the field and field-changes are measured. The field at the end of the first minute, for instance, is given by

$$F = 4\pi k(y_B - y_{B'})/A.$$

Two sudden interruptions in the record, such as are caused by lightning flashes, are shown at P and Q. At P the record suddenly touches the dotted line, indicating that for a moment the field was completely destroyed and the original induced charge ran to earth through the electrometer. At Q the movement extends beyond the dotted line, showing that the field was actually reversed in sign by the discharge and reached a reverse value equal to half its original strength. The shape of the record immediately after the passage of these lightning flashes, the recovery curve as it is called, is of considerable interest; it will be seen in these two examples that the thundercloud fields and charges recovered their original values after the lapse of something like 20 sec. The rate of recovery was approximately exponential, being most rapid immediately after the flash and becoming very slow as the cloud approached its original electrified condition. We may find the magnitude of these sudden field-changes by the same method used for the field itself; if Δy be the movement PP', or QQ', the sudden field-change which caused it is given by :

$$\Delta F = 4\pi k \,.\, \Delta y / A.$$

The test-plate arrangement, using a plate 50 cm. in diameter, is suitable for the measurement of fields of strength exceeding 1000 v./m., such as are caused by thunderstorms when they come within a distance of about 8 km. An inverted test-plate of smaller dimensions mounted on a pole and with a cover periodically and automatically swung below it for a short time has been used by Wormell for measurements of the large fields due to near storms. This is of great use during heavy rain. For more distant storms Wilson employed a different form of exposed conductor, a copper ball 1 ft. in diameter mounted on insulators at the end of an iron pipe 5 m. in length (Fig. 12). When the arrangement is in use the pipe is held vertical; by turning it about a horizontal axis through its base, the pipe is lowered and the ball enters an earthed metal case which shields it from the field.

Connection to the capillary electrometer is made by an insulated wire passing down the centre of the pipe.

The field producing a given movement of the electrometer is calculated as follows : The charge q induced upon the ball when raised to a height h above the ground must be such as to maintain it at earth potential, for the ball is earth-connected through the electrometer. Thus if V is the undisturbed potential of the air at the point occupied by the centre of the sphere of radius r, we have

$$V + q/r - q/2h = 0,$$

the third term allowing for the effect of the electrical image of the charged sphere in the earth. If the movement of the electrometer caused by raising the ball from its case to the height h is y, we have

$$q = -ky$$
and $V = k(1/r - 1/2h)y$.

This observation therefore gives for the mean potential gradient between the ground and the height h,

Fig. 12.—Elevated sphere for measuring fields of distant thunderstorms.

$$F = V/h = (k/h)(1/r - 1/2h)y.$$

The same equation relates the sudden changes of field ΔF to the corresponding sudden movements Δy of the electrometer. Since these relations, like those derived for the test-plate, are derived on the assumption of a perfectly plane earth, small corrections must be applied to allow for

the effect of the concentration of the lines of force upon the hut used to house the recording instrument.

The effects obtained from the elevated sphere become too small for accurate measurement when the storm is more than 20 km. away ; for still greater distances a wireless aerial has been used by some workers.   Such a wire can, however, give rise to unwanted corona (point-discharge) effects if the field is fairly big ; these are avoided by the use of the large sphere.   Thunderstorm fields are of the same order as the fine-weather field, about 100 v./m., when the cloud is 20 km. off.   At distances of 6 to 8 km. they often reach 5000 v./m., and when the storm passes overhead values of 10,000 to 20,000 v./m. are commonly observed, while the sudden changes may exceed these figures very considerably for a small fraction of a second.   In an active thunderstorm the fields are constantly altering as a result of the neutralisation of part or all of the charge by lightning discharges, of the subsequent rebuilding of the destroyed charges, and of movements of the cloud as a whole.

(b) *The amplifier-oscillograph method.* This was developed by Appleton, Watson-Watt and Herd [5] for the investigation of very rapid field-changes from distant storms (atmospherics) and has been used by them and others to record the details of field-changes which the capillary electrometer method, which has a time discrimination of about 0·1 sec., is unable to resolve.

The essential features of the arrangement used are shown in Fig. 13, in which C_0 is the capacity of an exposed conductor such as the ball of Fig. 12 (or for very distant storms, a horizontal wire aerial) and R_1 is a resistance of some 50,000 ohms inserted to render the aerial circuit aperiodic. C_1 is a condenser which is variable in steps to suit the field to be measured and R_2 is a high-resistance selected so that the time-constant $C_1 R_2$ is large compared with the duration of the field-change to be recorded.

If the field alters by ΔF as a result of a change ΔV of the potential at the effective height h of the exposed conductor, a change $-\Delta q$ must take place in the charge on

the conductor and the charge on the upper plate of C_1 must increase by $+\Delta q$.

This will produce a potential difference Δv across C_1 given by

$$\Delta v = \Delta q / C_1 = \Delta V - \Delta q / C_0,$$

since both these expressions give the change of potential at the point X.

Hence $\quad \Delta v = \dfrac{C_0}{C_0 + C_1} \cdot \Delta V = \dfrac{C_0}{C_0 + C_1} \cdot h\Delta F.$
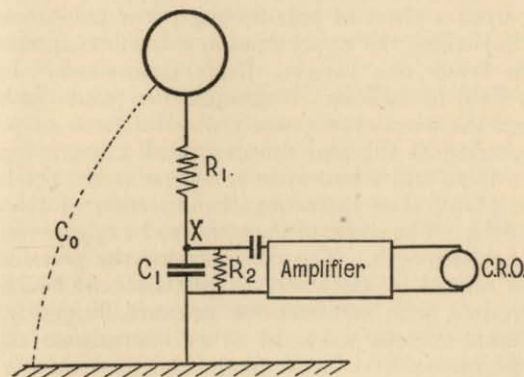
FIG. 13.—Input circuit for amplifier-oscillograph method (Appleton, Watson-Watt and Herd).

The arrangement thus acts as a condenser potentiometer and the amplified value of Δv causing the deflection of the oscillograph spot is proportional to the field-change. The oscillograph holds this deflection momentarily ; its deflection drifts back to zero in a time set by the quantity $C_1 R_2$ provided the time-constants of the amplifier are larger than this. $C_1 R_2$ can usually be made long enough for the faithful recording of all but comparatively slow changes of field.

If C_1 and R_1 are removed from the circuit, the arrangement of Fig. 13 can be used to record the rate of change of field, dF/dt. For then

$$h dF/dt = dv/dt + (dq/dt)/C_0 = dv/dt + v/C_0 R_2,$$

since the current i in the resistance R_2 is $v/R_2 = dq/dt$. If $C_0 R_2$ is small enough, the second term is much greater than the first and $v \simeq h C_0 R_2 dF/dt$.

(c) *The Alti-electrograph.* This instrument was devised by Simpson and Scrase [6] for attachment to a small balloon to measure directly the electric fields at points near and within active thunderclouds. Two separate iron pins resting upon a sheet of pole-finding paper are connected to vertical wires, the upper one on a bamboo framework and the lower one hanging freely downwards. In an electric field of sufficient magnitude for point discharge to charge the wires appropriately, the difference of potential appearing at the pins causes a small current to flow between them and a blue stain to appear at the positively charged point, thus indicating the direction of the prevailing field. The width of the stain can be related roughly to the field-strength. The disc carrying the pole-finding paper is rotated by clockwork to give a record which can be correlated with simultaneous pressure, humidity and temperature records made by other instruments carried on the flight.

(d) *The electric fluxmeter method.* The principle of this method has been described in § 17. When the device is carried in aircraft inside clouds, care has to be taken to eliminate the " autogenous " field produced by the charge collected by the aircraft itself. This is done (Ross Gunn) by employing two fluxmeters, one on the upper surface of a wing and the other, facing downwards, on the lower surface. For one of these the external and the autogenous field will be in the same direction and for the other they will be in opposite directions. The deformation of the field due to the aircraft itself can be corrected for by model experiments. Owing to icing difficulties this device cannot be used at levels where the temperature is much below 0° C.

A high-resolution form of this instrument devised by Malan and the writer has been employed in observations at ground-level to fill the gap between method (a) which has a poor time-resolution for field-changes and method (b) which is more suitable for rapid than for slow field-changes. Records of the output of the fluxmeter are made on a magnetic tape recording machine and those portions of interest are played back and recorded photographically later on, thus providing economical high-speed recording of transient field-changes.

(e) *The point-discharge galvanometer method.* If a fine point erected on a tall pole is insulated and connected to earth through a recording galvanometer, the current I due to point-discharge is roughly related to the prevailing field F by the equation $I = a(F^2 - M^2)$ where a and M are constants for the particular arrangement used. This method, which has been examined in detail by Whipple and Scrase,[8] provides a useful general record of the sign and approximate value of F for near storms but its time-resolving power is low and the above equation does not hold with much accuracy when the usual thunderstorm winds are blowing.

33. The Distribution of Electricity in Thunderclouds. All the methods described in § 32 have contributed to the solution of the problem of the distribution of electricity in the thundercloud. After some controversy it is now generally agreed that the main distribution is one in which an upper positive charge lies several kilometres above a lower negative one. This pattern may be repeated in different parts (cells) of a large cloud. The arguments of C. T. R. Wilson [2] on which this conclusion was first based, are set out below.

Consider a charge Q which occupies a spherical region within the cloud, its centre at height H above the ground.

The field at P (Fig. 14) is that due to an electric doublet of moment $2QH$, the opposite charge induced on the earth by Q being replaced by the image-charge $-Q$. This field is then $F = 2QH(H^2 + L^2)^{-3/2}$ and is directed vertically downwards if Q is a positive charge.

A lightning discharge involving Q may be of two kinds. If it is a discharge within the cloud, Q may most simply be considered to move vertically from a height H_2 to a height H_1, where it neutralises an opposite charge $-Q$. If it is a discharge to ground, Q disappears and F becomes zero. The net field-change produced by a cloud discharge is then

$$\Delta F_C = F_1 - F_2 = 2Q[H_1(H_1^2 + L^2)^{-3/2} \\ - H_2(H_2^2 + L^2)^{-3/2}] \quad . \quad (3.1)$$

and that due to a discharge to ground is

$$\Delta F_G = O - F = -2QH(H^2 + L^2)^{-3/2} \quad . \quad (3.2)$$
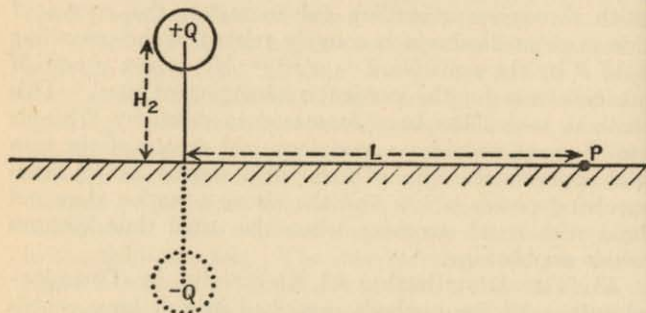
FIG. 14.—Thundercloud charge above the earth.

From observations of the sign of these field-changes one can infer the polarity of the cloud dipole responsible for them. If Q is a positive upper charge $(H_2 > H_1)$ eqn. (3.1) shows that ΔF_C will be positive for all values of L less than a critical value L_0 and negative for all values of L greater than L_0. L_0 will be given by

$$H_1(H_1^2 + L_0^2)^{-3/2} = H_2(H_2^2 + L_0^2)^{-3/2}.$$

Again, if Q is a negative lower charge moving from height H to ground, eqn. (3.2) shows that ΔF_G should be positive at all distances.

Observations of ΔF_C and ΔF_G at various distances have verified these deductions and hence indicate that the majority of thunderclouds carry positive charges above and

negative charges below. Table VI gives the results of observations of ΔF_G which show predominantly positive field-changes at all distances.

TABLE VI

SIGN OF FIELD-CHANGES DUE TO FLASHES TO GROUND AT ALL DISTANCES

Positive.	Negative.	Observer.	Place.
67	4	Schonland	Somerset East, S. Africa
267	16	Halliday	Johannesburg, S. Africa
77	13	Jensen	Nebraska, U.S.A.
411	33		

As concerns ΔF_C, eqn. (3.1) and (3.2) show that both cloud and ground flashes should produce positive field-changes when the storm is near but that the former should reverse to negative changes when the storm is at a distance greater than L_0. Since in storms in tropical and semi-tropical regions cloud discharges considerably outnumber ground discharges, we should expect to find in such storms a preponderance of negative field-changes at a distance and of positive ones when the cloud is near. That this is the case is shown by the data from South Africa in Table VII.

The South African observations in the Table offer strong evidence for the reversal effect. Those from Great Britain support them in showing a considerable preponderance of positive field-changes at small distances but in this case field-changes of either sign are about equal in number at distances exceeding 15 km. The reason for this is that in Britain cloud and ground discharges are approximately equally frequent whereas in South Africa the former considerably outnumber the latter.

Observations of actual zero values for ΔF_C at the reversal distance $(L = L_0)$ are rare. In practice the flash does not usually take a vertical path between H_2 and H_1. For this reason and because H_2 and H_1 vary from one flash

to another the concept of a reversal distance must be re-
placed by that of a reversal zone, in which small field-
changes of either sign are equally probable. This zone
may extend over a range of from 7 to 15 km. from the point
of observation.[4]

TABLE VII

SIGNS OF DISCHARGE FIELD-CHANGES OF ALL TYPES

$L_{kms.}$	ΔF + ve.	ΔF − ve.	Observer.	Country.
Distant Storms.				
> 15	250	2375	Schonland	S. Africa
> 10	113	644	Halliday	S. Africa
> 15	1699	1546	Wormell	England
Near Storms.				
< 5	54	16	C. T. R. Wilson	England
< 7	563	70	Schonland	S. Africa
< 6	86	18	Halliday	S. Africa
< 5	143	63	Wormell	England

Further strong evidence that the lower charge is usually
a negative one has been given by direct observations of
the polarity of lightning discharges to earthed metal
structures (§ 41). These have been made in Europe,
Russia, Japan and the United States and show that 90
per cent. or more of such flashes discharge negative
electricity. The total number of direct observations
available in 1941 was 2505 and 2422 indicated negative
discharges.

In conformity with the general picture of the charge
distribution outlined above, it is generally observed that
the electric field from a nearby storm just before it dis-

charges (the pre-discharge field) is negative, since the
lower pole is nearer to the point of observation, while at
a considerable distance it is found to be positive, the effect
of the upper pole being then predominant. In these
observations there is, however, a complicating factor of
some interest. It is frequently observed that just before
a very close discharge to ground the pre-discharge field
becomes small and may even become positive.

This effect is explicable if there exists below the lower
negative pole of the cloud an additional small " pocket "
of positive charge either in the base of the cloud or in the
air just below the base. Direct evidence for this pocket
is discussed in the next section. Indirect evidence is
afforded by the observation of Wormell[3] that there is a
type of cloud (or air) discharge which reverses in sign at
a much shorter distance than the normal and higher dis-
charge between the main poles of the cloud and which is
responsible for some of the apparently anomalous negative
values of ΔF from near discharges quoted in Table VII.
The type β leader process discussed in § 36 leads to the
same conclusion.

All these deductions as to the distribution of charge in
the thundercloud have been very beautifully confirmed
and extended by the alti-electrograph balloon observa-
tions of Simpson, Scrase and Robinson.[6, 9] These have
shown that in a balloon ascent near to or through the active
centres of a thundercloud, the field is frequently positive
on entering the cloud-base, after which it becomes negative
for about 1 km., then positive for as much as 4 km. and
thereafter it remains negative to the top of the cloud and
beyond. The results from a typical sounding of this type,
in which the width of the looped lines indicates the order
of magnitude of the field and negative and positive fields
are shown as open and hatched loops respectively, are
given in Fig. 15 (a). The figure also shows the dis-
tribution of the three charges, P, N and p which were
suggested to account for the observations. The values
found for the heights of these charges, and so for the
temperature levels at which they lie vary considerably from

storm to storm. It is not possible to be certain what path the balloon has followed in its fairly slow flight through the cloud nor whether the charge distribution derived from the results is that existing at the moment a flash takes place, but typical values for large thunderclouds in England are P, 7 km. (− 27° C.), N, 2·5 km. (− 4° C.) and p, 1·5 km. (+ 2° C.).
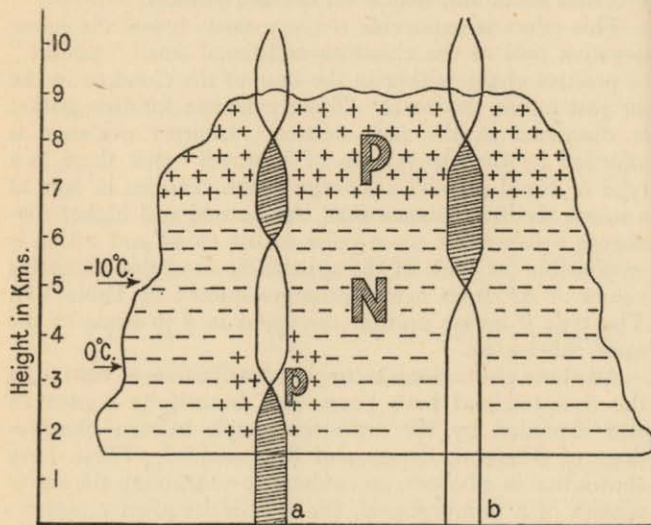
FIG. 15.—Typical alti-electrograph soundings (after Simpson, Scrase and Robinson)

The small p charge (which is called Q by Simpson, Scrase and Robinson) is not always observed, either because it is absent or because its small magnitude and dimensions cause the balloon to miss it. In such cases the record obtained is as in Fig. 15 (b).

Further confirmation of the general $P − N$ dipole distribution in thunderclouds has been obtained by Gish and Wait (§ 30) using the electric fluxmeter in aircraft flights above the tops of thunderclouds.

34. Magnitudes of the Electrical Quantities Involved in Thunderclouds.

When the distance L is large compared with the heights H_1 and H_2 of the main P and N cloud-charges, eqn. (3.1) takes the form

$$\Delta F_C = 2Q(H_1 - H_2)/L^3$$

or

$$2Q(H_2 - H_1) = - \Delta F_C \times L^3 \qquad . \qquad (3.3)$$

The quantity on the left is the difference between the electric moments of the thundercloud charges taking part in the discharge and is called the electric moment, M, of the discharge dipole. To determine it, the field-change due to a distant cloud flash and the horizontal distance L are required. The most frequent value of M determined in Great Britain [3] is 110 coulomb-km. though values as high as 1200 coulomb-km. have been observed. The average value for South Africa is 90 coulomb-km.[4]

Since $H_2 - H_1$ is approximately 2 km. the average quantity of electricity, Q, neutralised in a cloud discharge is about 20 coulombs. The range of variation of Q in individual discharges is from 2 to 200 coulombs if the vertical length of the discharge is supposed constant. If this length could be measured in every case, the variation would probably be much smaller.

The average electric moment of discharges to ground, which usually involve the lower negative pole only, is found to be of the same order as that of discharges within the cloud.

A single cloud-flash appears to discharge most of the section of the cloud which is producing lightning at the time. A fairly active thunderstorm produces one flash every 20 sec. To feed these flashes alone, it would have to separate out charges of opposite sign at an average rate of 1 coulomb/sec. and the generating mechanism would have to provide an average current of 1 amp. Reference has been made in § 32 to the rapid rate at which the initial regeneration of the field takes place after a flash. In the case of cloud discharges it is usually such as to be able to restore the whole field in about 5 sec. if it continued for that time. The exponential form of the recovery curves

suggests that this initial rate of supply of charge remains constant but that an opposite neutralising effect occurs which tends to diminish the rate of increase of the charges and which is proportional to the quantity of charge present at any time.

That neutralising or dissipating effects of this nature exist below the cloud is shown by measurements of the mean current from positive point-discharge flowing upwards from the earth to the N charge (§ 29). This current is of the order of 2 amp. A similar dissipating current bringing negative charge down from the upper air to neutralise the P charge has been observed (§ 30). Its average value is about 1 amp. but it has been found on occasion to reach 6·5 amp.

The electrical energy generated by the cloud is thus only partially spent in the production of lightning discharges; just before a discharge most of the power of the generating mechanism is being used to overcome leakage effects opposing the building up of the main dipole. It may happen that the field, owing to this leakage load, never reaches the sparking value and the cloud, though electrified, does not produce lightning.

An estimate of the maximum potential reached by the two main cloud charges can be found from an argument given by C. T. R. Wilson.[10] Consider first the simple case of a charge Q contained within a sphere of radius R. The electric field at the surface is then given by $F = Q/R^2$. The maximum value of F for a discharge to occur is known from experiments by Macky on the behaviour of water drops in strong fields to be about 10,000 v./cm. or 33 e.s.u. This figure is one-third of that in normal air; the reduction is due to the fact that in strong electric fields a water-drop becomes unstable and pointed filaments are drawn out from opposite ends. For an average value of Q of 20 coulombs or 6×10^{10} e.s.u. the above equation gives $R = 427$ m. and the dimensions of the charged region are of the order of 1 km. The potential at the surface of the sphere is given by

$$V = Q/R = FR = 4\cdot 2 \times 10^8 \text{ volts.}$$

If the radial electric force everywhere within the sphere were such as to produce sparking, the potential at the centre would be $2FR$ or about 8×10^8 v. Two such spheres carrying opposite charges would therefore be at a difference of potential of $1\cdot 6 \times 10^9$ v. A similar result follows from the very different assumption that the two poles of the cloud are distributed in horizontal layers with a neutral region of thickness D, between them. If F is 10,000 v./cm. and D is 2 km., the difference of potential between them is 4×10^9 v.

Recent work on the manner of development of the lightning discharge (Chapter IV) suggests, however, that the average value of the field at the surfaces of the charged regions is considerably lower than 10,000 v./cm. Once a short leader streamer (§ 36) has been produced in a limited region where the field is momentarily 10,000 v./cm., its further development may be possible in fields which are only one-tenth as large. Actual measurements of fields inside thunderclouds by Ross Gunn [7] support this conclusion, for the highest " undistorted field " value reported was 1600 v./cm. just before the observing aircraft was struck by a discharge and the average maximum field for nine different thunderclouds was 1300 v./cm. If F is one-ninth of the value used above, the diameter of the spherical charged region ($2R$) becomes about 2·5 km., the potential reached before discharge $1\cdot 4 \times 10^8$ v. and that between the poles $2\cdot 8 \times 10^8$ v. These are likely to be lower limits for the potentials.

The disappearance of 20 coulombs in a lightning discharge is thus estimated to involve the dissipation of at least 6×10^{16} ergs or $1\cdot 6 \times 10^9$ calories, most of which is converted into heat along the path of the lightning flash. A cloud giving one flash every 20 sec. is dissipating electrical energy in the form of lightning at an average continuous rate of at least 3×10^5 kw. To maintain in addition currents of the order of 2 amp. above and below it involves the provision of a further 6×10^5 kw.

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CHAPTER IV

THE LIGHTNING DISCHARGE

35. General Features. The most frequent form of lightning is the *cloud discharge*. Though most cloud flashes are known from field-change records to be discharged between the main P and N charges of the cloud dipole (Fig. 15), there is evidence that others pass between the N charge and a lower positive charge, either the small p pocket of Fig. 15 (*N-p discharge*) or a positive space-charge in the air below the cloud (*air-discharge*).

A similar type of flash between the P charge and negative charge drawn down from the upper air to the top of the cloud or to the air immediately above it, would account for many cloud-flashes which are seen at night to be restricted to the cumulus heads of thunderclouds and the regions above them.

Most ground-discharges, as has been shown in § 33, remove negative charge from the N region. There is a good deal of evidence to show that they are usually triggered off by the strong field between the N and p charges or the positive space-charge in the air, thus extending an $N-p$ or an air discharge all the way to the ground. The first strokes of such discharges thus bring to ground the residue of the negative charge left after the smaller and lower positive charge has been neutralised.

A general account of the various forms of the lightning discharge and their effects will be found in reference 1.

Many lightning flashes, both within the cloud and to ground, are observed to flicker. Early camera studies by Walter and others showed that this is due to the fact that the discharge consists of a number of separate components or strokes. Though the whole flash often occupies more than a second, the luminosity of the component strokes is generally over in a few milliseconds. After each stroke the ionisation along its channel, though much reduced by

electron-attachment and by recombination and diffusion of the ions, usually persists in sufficient measure to provide the following stroke with a ready-made path which it follows. If, however, the time-interval is exceptionally long, the channel ionisation is too much weakened and the following stroke may take a different path.

The channel is often shifted several metres by the wind in the intervals between strokes ; the discontinuous nature of the discharge is then shown as *ribbon lightning*.

Observations with moving cameras show that the branching of the lightning discharge to ground is almost always confined to the first stroke of a series.

The maximum number of strokes so far observed from a single flash to ground is 42. The most frequent number is 3 or 4 and single-stroke discharges are quite common. The time-interval between strokes is variable, the most frequent value lying between 0·03 and 0·07 sec.

36. Luminous Streamer Processes in the Discharge.

(a) *Experimental methods.* To investigate the nature of the luminous processes in the lightning discharge much use has been made of a method of high-speed photographic recording devised by Sir Charles Boys. If a camera is so constructed as to provide relative motion of lens and film, the recorded image of any series of luminous events in two-dimensional space will be distorted in such a way that later events are progressively displaced in position relative to earlier ones. Comparison of the distorted picture of a lightning discharge with one taken on a fixed camera would then show the order in which light was emitted at different parts of the channel. In practice, however, this is very difficult unless the distortion is large, because both fixed and moving cameras require an accurately placed reference line for the comparison. The difficulty is avoided in the Boys camera by the use of two lenses rotating in a circle ; these give two pictures with distortions in opposite directions and a reference line provided by a diameter of the circle. This principle has been used to follow the processes in the discharge with a time resolution of 1 microsec. Full details of the Boys

camera and of later forms of camera which have been used by the writer and his associates in South Africa are given in references 1 and 2.

(b) *General results.*[1, 3] Each separate stroke of a discharge to ground has been found to be of a dual nature. A downward-moving luminous process, the *leader* streamer, is followed upon arrival near the surface of the ground by a much more rapid and intensely luminous main or *return* streamer which re-traverses the channel in the reverse direction. The leader to the first stroke is of a different nature to the leaders to subsequent strokes ; it takes about ten times as long to reach the ground and its luminosity is intermittent instead of continuous.

Fig. 16 (b) (after Schonland, Malan and Collens) illustrates these points in the case of a discharge of three strokes which has been photographed by a single-lens camera in which the lens has moved in a straight line from left to right. It is actually a picture in two dimensions of space and one of time since horizontal displacements with reference to the fixed picture (a) represent time as measured from the point P. These displacements are not shown on a uniform scale in the diagram. The Boys camera gives two simultaneous pictures from two lenses moving in opposite directions and thus avoids the use of a fixed picture.

The figure shows that the leader to the first stroke proceeds in a series of steps. Each of these is about 50 m. in length (range 10 to 206 m.) with fairly regular pauses (31 to 91 microsec.) between the steps. After each pause there is almost always a fresh direction of travel, and these changes in direction are responsible for the tortuous form of the lightning channel. The branches of the leader, which are usually found in the first stroke only, are similarly developed in steps and move outwards from the main channel. Each step is the bright tip of a long streamer stretching down from the cloud, the stem of which is faintly illuminated.

When the most advanced branch of the stepped leader system approaches within from 15 to 50 m. of the ground,
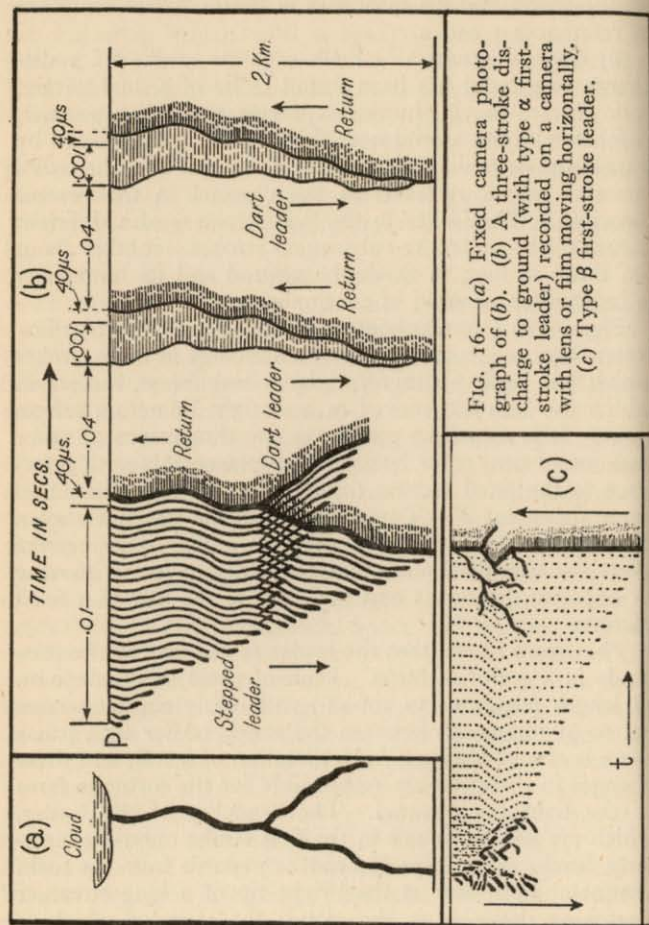
FIG. 16.—(a) Fixed camera photograph of (b). (b) A three-stroke discharge to ground (with type α and first-stroke leader) recorded on a camera with lens or film moving horizontally. (c) Type β first-stroke leader.

the return streamer jumps the gap and passes rapidly back from ground to cloud, running outwards along each branch. This process takes microseconds to traverse parts of the channel originally created in milliseconds by the stepped leader.

Stepped leaders are of two types. In the type α leader (Fig. 16 (b)) the steps are short and are often difficult to photograph because they are not very bright. In the type β leader (Fig. 16 (c)) the upper part of the channel is formed by longer and brighter steps than the lower portion which has very weak type α steps. This difference is associated with extensive branching in the upper part of the channel. It suggests that there is a stronger electric field, due to positive space-charge in the air below the cloud, at the top of the leader path than later on. Electric field studies of the leader process show that although photographic records of events in the air below the cloud indicate that only 10 per cent. of leaders are of type β, this type is actually the most frequent one. The β stage in the leader is usually hidden within the cloud, the high field regions required to produce it presumably lying between the N and p charges of Fig. 15.

In the case of strokes subsequent to the first, the leader is not stepped but travels continuously to ground. As shown in the figure, it records as a continuous bright line followed (in the time dimension) by fainter illumination. This indicates that the leader takes the form of a fast-moving javelin, or dart, with a bright point about 40 m. long and a long, faintly-illuminated stem extending back to the cloud. For this reason it is termed a *dart* leader. The only exceptions to the continuous movement occur when an unusually long time-interval has elapsed since the preceding stroke ended. The dart leader is then much slower; frequently it shows the stepped method of progression over part of its length and takes a path different from that of its predecessors.

A typical first-stroke return streamer record is shown in Fig. 17. Times are shown in microseconds after the start at O.

The streamer covers the distance of about 2 km. between ground and cloud in 60 microsec. The figure shows that
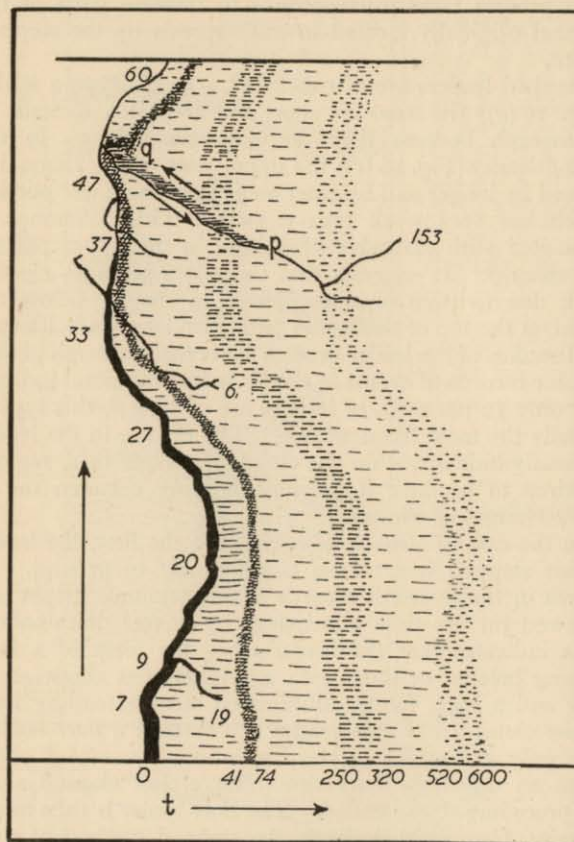
FIG. 17.—Return streamer time-table (Malan and Collens). (Times in μ-secs.)

it continues to remove charge from the cloud for a further 600 microsec. The brightness and velocity of the streamer decrease as it travels upwards but are revived suddenly

as each of the branching points is reached, indicating, as has been shown by the detailed studies of Malan and Collens,[4] that the charge on a branch calls into existence a fresh flow of current from the ground. The current at the base of the channel is subject to violent fluctuations (*M* components) from these effects. Further fluctuations, (like the last two in Fig. 17), are produced by processes which are hidden in the cloud. In the case of subsequent strokes these latter fluctuations are the only ones observed because there are usually no branches to be traced by the return streamer.

The average velocities of the three processes which have been described, the dart, return and step streamers, are 2,000, 20,000 and more than 10,000 km./sec. respectively. Although individual steps are pushed forward at > 10,000 km./sec., the step-streamer process as a whole, as Fig. 16 shows, is slow, for its numerous pauses cause it to have an *effective* velocity which seldom exceeds 300 km./sec. and can fall as low as 100 km./sec.

In the case of very tall structures like the Empire State Building, 1250 ft. high, McEachron and Hagenguth[5] have shown that the lightning discharge frequently begins with an *upwardly* directed stepped leader which travels from ground to cloud and is not followed by a return process. The current continues to flow for some time in the channel thus formed, after which a normal downward dart leader, followed by an upward return streamer, creates the second and later strokes.

37. Electrostatic Field-changes associated with Lightning Processes. The processes described in § 36 give rise to electrical field-changes which are best studied by the amplifier-oscillograph method of § 32 (*d*) or the electric fluxmeter of § 17.

Fig. 18 (A) shows the general form of the field-changes produced by a ground discharge of three strokes when near (3-5 km.), but not directly over, the recording instrument.[6, 7] Fig. 18 (B) shows the form taken by the same field-changes when the flash is fairly distant (12-20 km.). In this case, as described in § 40, induction

and radiation field-changes would be superimposed on the electrostatic field record but for the sake of clarity they have been omitted from Fig. 18 (B).

The manner in which various sections of the field-change record have been identified with the various streamer-processes observed with the Boys camera is shown in the figure. The *l* portions of the record correspond to the downward movement of leader streamers and the two *r*
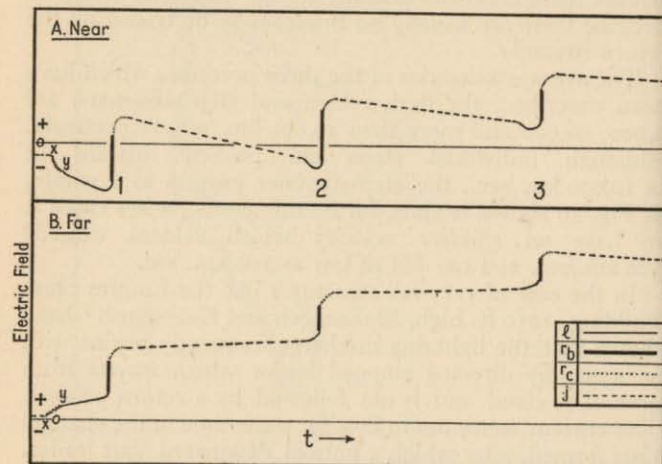
FIG. 18.—Electrostatic field-changes due to a discharge to ground (Schonland and Malan).

portions, r_b and r_c, with the upward return streamer. The remaining *j* portion corresponds to a streamer process which cannot be photographed since it takes place within the cloud. From a detailed study of records of this kind the following conclusions can be drawn :

(1) *Leader* (*l*) *field-changes*. These, which produce hook-shaped negative changes when nearby and parabolic positive changes when at a distance, indicate clearly the lowering of negative charge towards the ground by the leader streamers. Since they remove charge from an

initial height, H_2, to a lower effective height, H_1 (§ 33, eqn. 3.1) when the charge is distributed along the channel, their net field-changes reverse in sign with distance.

The leader to the first stroke in Fig. 18 shows a field-change which has a much larger slope over the first section *xy* than later. This is because it is a type β leader (§ 36 *b*), and lowers charge more rapidly in its initial than in its final stage.

As Fig. 18 shows, dart leaders to subsequent strokes occupy a much shorter time than the first stepped leader but show a similar form of field-change.

It is of some importance that the durations of leader field-changes are often much longer than the durations of the corresponding leader processes recorded by cameras operating at the same time. For this implies that a considerable length of the leader channel, sometimes of the same order as the visible length below the cloud, is hidden inside it and that the *N* charge creating the leader is at a considerable height.

(2) *Return streamer* (r_b and r_c) *field-changes*. Since these involve the removal to ground of the negative charge on the leader (r_b) and remaining in the cloud (r_c) their field-changes are positive at all distances. The rapid r_b change occupies a time of the order of 50 microsec. and has a complex fine structure which is not shown in Fig. 18. The slower r_c change, which occupies 1 or 2 millisec., is considered to be due not only to the final mopping-up of the charge remaining in the region which produced the leader, but also to an upward flow of positive charge induced on the top of the channel by *N* regions as yet untapped by the discharge.

(3) *Junction* (*j*) *field-changes*. These changes occupy the intervals between strokes. Since they are negative when near and positive when far, they must be created by slow movements of charge within the cloud, in which positive charge moves upwards or negative charge moves downwards. There is some evidence [7] that in most cases they are due to the upward movement of positive streamers

from the top of the channel to new regions of negative charge, which they render conducting.

(4) *The mechanism responsible for successive strokes.* It has been shown by Malan and the writer [7] that a number of lines of evidence leads to the conclusion that each stroke of a discharge to ground comes from an N region which is higher than that producing its predecessor. From this and other information it is deduced that the N regions creating a discharge of many strokes are not haphazardly distributed within the cloud but form a single and approximately vertical column which is intermittently discharged and which extends upwards to considerable heights. A ground flash of many strokes may come from a column 6 km. long. These authors have shown that the pauses between strokes represent the time required for the junction-streamer process to move upwards and render the next section of the charged column sufficiently conducting for it to discharge down the channel as a leader.

38. The Mechanism of Streamer Development.[8]

It is known from laboratory studies of gas-discharge processes that an electrical streamer possesses a conducting stem filled with ions and electrons and that the motion of these charged particles in the field prevailing inside the stem constitutes a current which keeps the streamer surface, and its tip in particular, supplied with charge as it moves forward. For forward movement the electric field at the tip must be sufficiently great to cause free electrons in the air in front of the advancing streamer, some of which are created photoelectrically by ultraviolet light from the streamer itself, to ionise by collision. This ionisation is cumulative and leads to the production of electron avalanches which extend the streamer tip.

The streamers give out light at and behind their tips because excitation as well as ionisation occurs in the strong field region. Since the excited atoms and molecules return a short time later to their ground-states, and since the tip moves forward during this interval, the region of intense luminosity extends backwards for a limited distance behind the tip.

In the case of the dart leader, the negatively charged streamer is guided along a pre-existing channel because ionisation created by previous strokes exists in front of the advancing tip. It can be shown that the velocity attained by the streamer in this case should be proportional to the cube root of the pre-existing electron density, n, in front of it.[8] When the interval between strokes is short, n is large and high dart-leader velocities are possible, but when the interval is long and n has been reduced by electron attachment to atoms and molecules and by diffusion processes, the dart leader velocity is low.

The fast return-streamer also travels along a previously ionised channel, more heavily ionised and fresher than that available to the dart leader. In this case, however, as was first shown by Simpson, the tip of the streamer is positively charged and the electron avalanches in front of it move inwards in a strongly converging field. These factors account for the high velocity of the return streamer.

The stepped leader, which is the first of all the discharge processes, appears from the photographic records to involve advance into virgin air and to follow no pre-existing trail of ionisation. Its slow effective velocity, which is of the same order as the minimum velocity at which an electron can ionise by collision, coupled with the very high velocity of the step streamers themselves suggests, however, that it too follows an ionised path prepared by an unrecorded *pilot* streamer, or a cone of such streamers, which precedes it. On this view the steps are secondary effects superimposed upon the pilot streamer and catching up at intervals with it.

Such pilot streamers would be controlled in direction by purely local variations in the electric field in front of their tips and their paths will be tortuous and branching, especially if the air contains concentrations of positive ion space-charge.

A fuller discussion of these streamers and of the unsettled question of the mechanism responsible for the fairly regular pauses observed in the stepped process will be found in references 9, 10 and 11. It is generally agreed

that these pauses owe their origin to a property of the streamer itself [8] and not to any intermittent feature in the discharge from the cloud but whether this property involves a periodic transition from a " glow " to an " arc " form of streamer or is simply due to the streamer overreaching itself and requiring a pause to renew the field-strength and the conductivity of its stem is not yet certain.

39. Discharges which do not Pass to Ground. Discharges which do not reach the ground, whether they are air-discharges or take place entirely within the cloud are found to be of the leader type, without the usual return process. Air discharges when photographed show complex branched type β step-leader streamers with dart leaders superimposed upon them. They travel towards local concentrations of positive space-charge below the cloud and weak " recoil " streamers moving back from such concentrations have been observed. An example of recoil is shown along p, q in Fig. 17.

The study of discharges which are entirely within the cloud has so far been possible by field-change methods only. These show that cloud discharges are of the same character as air discharges and can occupy times of the order of milliseconds.

40. Induction and Radiation Field-changes. The form of the electrostatic field-changes associated with lightning discharge processes has already been described. The complete electric field from a small dipole of moment M is actually the sum of the components represented by

$$E = M/r^3 + (dM/dt)/cr^2 + (d^2M/dt^2)/c^2r,$$

expressed in e.s.u. where c is the velocity of light and r the radius vector to the centre of the dipole. The first term corresponds with eqn. (3.1) for the electrostatic field, the second is the induction or magnetic term and the third the well-known expression for radiation from a Herzian dipole. Within 10 km. of a lightning flash the first term is much larger than the other two and the field-changes caused by lightning produce mainly electrostatic effects. At 25 km., however, the last term shows up strongly and at 50 km. and beyond, it is predominant.

The radiation field is proportional to d^2M/dt^2 and so to the rate of change of the current in the portion of the discharge concerned at time $t - r/c$. Radiation effects from lightning are therefore most prominently associated with processes which involve the rapid starting and stopping of streamers carrying large currents. These are (a) the stepped processes found in first stroke leaders and in cloud and air discharges and (b) the return processes in each stroke.

The radiation effects of these processes are found to constitute the majority of atmospheric (static) disturbances on the longer wavelengths used in radio communication and can be observed more than 4000 miles away from the parent discharge. They have been of practical value in the location of thunderstorms for meteorological purposes (sferics) and the manner in which they are modified by successive reflections from the ionosphere has been used to give information about the height and nature of its lower D layer.[1,12]

41. Currents and Current Variations. Much work has been done to determine directly the magnitudes of lightning currents to ground and their time-variations.[9,10] One instrument used for determining peak currents through elevated conductors and towers struck by lightning is a simple recording voltmeter, known as a Klydonograph, placed across one of a string of insulators acting as a potential divider and connected between the conductor and the earth. This device makes use of the fact that the diameter of the " Lichtenberg " figure produced around a positive point resting on a flat photographic film backed by an earthed plate is proportional over a certain range (up to 25,000 v.) to the potential difference between point and plate. To record discharges of both polarities two instruments, oppositely connected, are necessary.

The Klydonograph method requires a knowledge of the surge impedance between the conductor and the earth and this is not always easy to determine. More use has recently been made of " magnetic links," two or three bundles of short unmagnetised steel strips of high

retentivity which are mounted at different known distances from the conductor. The peak currents passing in the discharge can then be found from the remanent magnetism in the links, calibration being effected with known currents.

In a development of this device called the Fulchronograph (Wagner and McCann)[14] the links are carried on an aluminium wheel which rotates between coils through which the lightning discharge passes. The remanent magnetism of successive links then gives the time variation of the current.

The time variation has also been determined by shunting special types of oscillograph across resistances of high current capacity. The data are, however, as yet insufficient for a generalised statement of the manner in which the current in a discharge varies with time in the successive strokes of a discharge. In first strokes it rises to a peak of average value 20,000 amp. (maximum observed value 250,000 amp.) in from 1 to 20 microsec., with a most frequent "time to peak" of 2 microsec. Its manner of decay is variable but it is usually stated to fall to half its peak value in about 24 microsec. (range 7 to 115 microsec.).

Continuing currents of considerable magnitude and of long duration have been found by McEachron[13] to follow strokes to the Empire State Building but in discharges to lower structures the current is usually found to have fallen to very low values ($< 0 \cdot 1$ amp.) a few milliseconds after the start of the return portion of the stroke.[14]

The quantities of electricity found by integrating the current-time curves of direct discharges are of the same order of magnitude (average 30 coulombs, maximum 164 coulombs) as those determined by the method of C. T. R. Wilson (§ 34).

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CHAPTER V

CHARGE GENERATION IN CLOUDS

42. Introduction. Many suggestions have been made as to the manner in which electric fields can be built up in clouds. Some of these have had to be discarded as facts on which to test them have accumulated. Most of the mechanisms left in the field may well be in operation at the same time and it is not easy to decide which are predominant in creating the electrification of thunderstorms.

An acceptable theory must (a) offer an explanation of polarity of the main P-N dipole ; (b) give P and N charges each of about 30 coulombs under reasonable conditions of temperature, water-content and updraught ; (c) operate at a temperature at least as low as $- 10°$ C., and (d) be shown experimentally to work. Subsidiary mechanisms may be necessary to account for the lower p charge and for observations to be described in § 45 on the relation of the charge on thunderstorm rain to the prevailing field.

Of the theories which have been rejected because they did not meet all these conditions the best-known is that of Simpson, which proposed to account for the main electrification of the cloud by the Lenard effect, the breaking-up of water-drops in ascending air-currents. This mechanism, which left the larger fragments with a positive charge can satisfy neither the temperature condition (c) above nor condition (a), since it requires the lower portion of the cloud to be positively charged. It may, however, as Simpson and Scrase have suggested, be responsible for the small p charge of Fig. 15.

43. The Separation of the Opposite Charges once they are Formed. All the mechanisms which are at present under consideration satisfy condition (a) because they all suggest two entirely different types of carrier for positive and for negative charge and because all give a

negative charge to the heavier of the two types. Polarisation of the cloud in the requisite direction then follows automatically by the action of gravity in pulling the negative charges down (Fig. 19). When an updraught in the cloud is capable of supporting or even of driving upwards the heavier negative group of carriers (raindrops or snow or large ice-particles) the lighter ones (droplets or ice-fragments) are driven upwards still faster, so that the
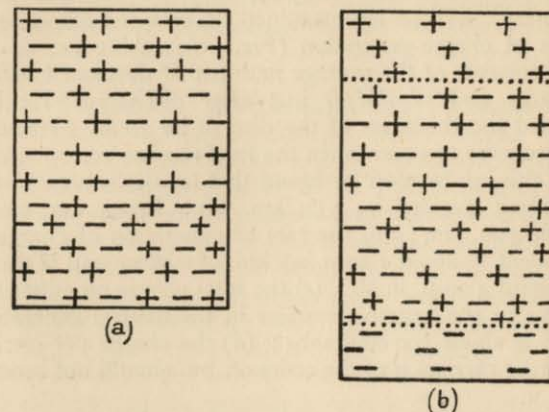
FIG. 19.—Gravitational separation of opposite charges (C. T. R. Wilson).

relative velocity of separation of the two groups is the same as if no updraught were present.

The shape of the field-changes immediately after a cloud-discharge (§ 32) suggests that the process of separation of the oppositely-charged carriers is one involving initially free fall under gravity and that this free fall is progressively retarded and finally reduced to nearly zero by the attractive forces between the separating charges. The rate of fall of the heavier carriers in this final stage must be considered just sufficient to maintain the charges constant in the face of dissipation currents within the cloud and directed to it from outside. Ultimately the field

between the separated charges may become strong enough for a discharge to pass. If the charges are equal, this leaves the cloud in its initial state and the larger carriers in the neutral region of Fig. 19 once more separate out under the action of gravity. The relative velocity of separation has been taken by C. T. R. Wilson[1] to be of the order of 6 m./sec., the terminal velocity of a water-drop 0·15 cm. in radius. Relative to such a drop, the much slower positive droplets may be considered as stationary. Wilson has examined the case of a disc-shaped region of charge-generation (Fig. 19) making use of his measurements of the average moment M developed before discharge and of dM/dt just after discharge. He has assumed the thickness of the disc to be about 1 km. and the discharge to occur when the field reaches 10,000 v./cm. From this information he found that (a) the volume of the generating region is 4 cu. km. and hence its cross-section 4 sq. km. ; (b) the rate of generation of charge is equivalent to about 1 amp./sq. km. of surface and is therefore about 4 amp. in all ; (c) the total charge on either the positive or the negative carriers in the neutral generating region is about 500 coulombs ; (d) the charge per c.c. on the larger carriers is of the order of, but should not exceed, 30 e.s.u.

The values deduced in this way are all in conformity with what is known of the quantities concerned. They are not seriously changed if modifications are made in the postulates to bring these into closer agreement with more recent information (thickness 3 km., breakdown field 1000-10,000 v./cm.).

A different type of cloud polarisation by gravity has been discussed by Frenkel in connection with fog-like clouds of small droplets. These are considered by him to be capable of capturing negative ions from the air (§ 44 (b)) which is then left with an excess positive charge.[2] The charged droplets are not held up completely by the internal field but drift slowly downwards at such a rate that the polarisation charge developed per unit area is equal to the charge neutralised by the internal conduction or dissipation

current in the same time. The internal field, E_s, in this "stationary" condition is related to the equilibrium field, E_q in which the drops are fully supported, by the equation $E_s/E_q = \Delta/\lambda$ where λ is the ionic conductivity of the air and Δ is the much smaller contribution to the conductivity provided by the droplets. E_s is found to be of the order of 100 v./cm., a hundred times less than Eq. In clouds where the water content is greater, the positive ions will attach themselves to larger particles and the development of polarisation in the thundercloud then appears to become the same as that considered by C. T. R. Wilson.

44. Suggested Mechanisms of Thundercloud Electrification. (a) *Mechanisms which depend on the selective capture of negative ions by the larger carriers.* A mechanism suggested by C. T. R. Wilson depends upon the selective capture of negative ions from the air by large falling water-drops or ice-particles, the residual excess positive charge in the air being absorbed by very small droplets or ice-fragments in the upper part of the cloud.

For the mechanism to work, the ions concerned must be large ones (§ 1). Since their mobility is about 0·0003 cm./sec., the velocity with which they move under the influence of the prevailing electric field is much less than the velocity of fall of the larger carriers.

Consider a large drop or ice-particle inside a cloud in which a downwardly-directed field already exists. Owing to this field the drop will have a negative induced charge on its upper surface and a positive charge below (Fig. 20). In the case of an ice-particle, Chalmers has shown that dielectric polarisation will produce a similar distribution of charge.[3] As these larger carriers move downwards, they will approach positive ions moving down more slowly and negative ions moving up ; encounters with other drops or ice-particles are considered to be relatively rare. The drops will behave differently in encounters with the two ion-streams. At the under-surface of a drop the negative ions will be attracted and the positive ions repelled and the former, if within a suitable distance, will be captured, giving

the drop a net negative charge. The negatively-charged upper surface of the drop does not reverse the process, because here it is receding from the ion-streams and positive ions have been repelled away from the drop in the course of their passage past it. Innumerable episodes of this kind can cause the drop to acquire a considerable net negative charge, without appreciably altering the chance of further captures. A water-drop 1 mm. in radius, falling in a field of 10,000 v./cm. carries at each end an induced charge of 0·25 e.s.u. If it has collected a net negative charge equivalent to the very reasonable value of 12 e.s.u./c.c., the increase in the upper negative charge and the decrease in the lower positive charge will be 10 per cent. only. Whipple and Chalmers have shown that the maximum charge which can be collected in this way if ions of both signs are in equal concentrations is $-0·52Fr^2$ when F is the field-strength and r is the radius of the drop. This gives 40 e.s.u./c.c. for the case just considered. If negative ions only are present the maximum charge collected is $-3Fr^2$, six times as great.

The tiny droplets and ice and snow particles, which fall relative to the air with a velocity of the same order as that of the ion-streams, will be unable to exercise the selective action described. They will therefore absorb a net positive charge, the complement of the negative charge collected by the larger carriers.

FIG. 20.—Capture of slow negative ions by large falling drops.

Wilson's mechanism operates in such a way as to enhance enormously a pre-existing field. This could be the normal field of fine weather or a field produced initially by one of the other mechanisms to be described. The Wilson process has been shown by Gott to work under laboratory conditions but it is not generally agreed that it is the chief agency in building up the fields of thunderclouds. One criticism which has been raised against it is that the concentration of large ions necessary for it to be effective is not likely to be present until the electric field is very strong.

The theory of Frenkel [5] is based upon a selective adsorption of small negative ions by water-drops or wet ice-crystals. Owing to the polar character of the water molecule, the skin of such carriers is considered to consist of an electrical double-layer with positive poles outwards. This tends to capture negative ions from the air until a difference of potential v_0 is created between the drop and the cloud of positive ions which surround it. The equilibrium negative charge acquired in this way is v_0r where r is the radius of the drop and v_0 is 0·25 v. This mechanism gives a considerable negative charge density to small drops but is not very effective in the case of large ones ($r = 0·1$ cm., charge density 0·02 e.s.u./c.c.) nor does it give very large fields. To carry it further, Frenkel considers that adjacent downward-moving droplets coalesce into larger drops as a result of hydrodynamical forces between them. These larger drops in their downward flight sweep up and collect the charge from slower-moving droplets, obtaining in this way a charge which varies as r^3 and creating much larger fields. The theory does not appear to have been tested experimentally.

(b) *Mechanisms which depend upon electrification as a result of change of state.* In its earliest stage of development the cumulonimbus cloud consists almost entirely of droplets of water. Some of the droplets in the portion of the cloud above the freezing level remain unchanged in state and supercooled. As the cloud top rises, an increasing proportion of these droplets are converted into ice-particles,

either directly or by the action of special "sublimation-nuclei". The effectiveness of these processes increases with decreasing temperature and hence with increasing height. Ice-particles formed in this way grow very rapidly at the expense of surrounding supercooled drops if the temperature is low. As these grow in size they fall and the mature thundercloud contains a complicated mixture of hard and soft hail, snow particles, rime-covered ice (graupel) and water drops. All these hydrometeors are in various stages of growth, or of disappearance, depending upon the temperature-levels to which they have risen or fallen and all of them are constantly colliding and coalescing with one another. The changes of state, and the collision and coalescence processes thus involved, offer a variety of opportunities for the generation of electric charge and a number of possible mechanisms have been based upon them, most of which have been tested experimentally. These are listed in Table VIII below.

Mechanism No. 7 is not one of change of state but is included in the Table because it involves ice ; experimental test of it has so far yielded conflicting results. Mechanism No. 6 would not create a dipole of the correct polarity and could only operate in the lower levels of the cloud where like the Simpson breaking-drop mechanism, it may contribute to the formation of the lower p charge and the strengthening of the N charge above it (Fig. 15). Experiments by Kramer indicate that mechanisms Nos. 2 and 3 of Findeisen must be regarded as of minor significance compared with No. 1.

The remaining mechanisms, Nos. 1, 4 and 5, are all based upon electrification resulting from the freezing of supercooled drops, either directly (5) or by contact with falling ice-pellets (1 and 4). If one or other of them is responsible for the formation of the main cloud dipole, the first occurrence of lightning should be closely related to the first development of the ice-phase in the cloud. This in turn is bound up with the height and so with the temperature-level to which supercooled water-drops are raised by the updraught within the cloud.

TABLE VIII

	Proposer.	Charge Carriers. Heavy.	Charge Carriers. Light.	Generating Process.	Sign of Charge on Heavy Carrier.	Remarks.
1	Findeisen and Kramer	Large rime-coated ice-particles	Small ditto	Coalescence of supercooled drops with ice to form rime	−	Small rime fragments +vely charged
2	Findeisen	Ice	Ice-splinters	Growth of ice by sublimation from drops	+	Found experimentally to be of minor significance (Kramer)
3	Findeisen	Ice	Ice-splinters	Evaporation of ice	+	As for (2) above
4	Workman and Reynolds	Ice	Fragments of supercooled drops	Glaze-ice formation on hail	−	Effectiveness increased by traces of $CaH_2(CO_3)_2$ in air
5	Chalmers	Ice	Water-drops	Sudden freezing of super-cooled drops	− ?	Not tested experimentally
6	Dinger and Ross Gunn	Water	?	Melting of ice at low levels	+	Presence of dissolved gases necessary before freezing if effect is to be appreciable.
7	Simpson and Scrase	Ice	?	Frictional electrification by collision	−	Inferred from Antarctic blizzard observations

Evidence as to the height of the larger droplets within the cloud at various stages has recently been made available by microwave radar studies of the so-called " cloud echo ". These suggest that full-scale electrification leading to fields sufficiently great to produce lightning does not occur until the supercooled droplets have been raised to a temperature level, variously given as between $-28°$ C. and $-40°$ C., where ice-crystal formation on an appreciable scale can occur.

These observations, together with the conclusions of Malan and the author (§ 37 (4)) that the N charge occupies a column of considerable vertical extent, whose top reaches to very low temperature-levels, make it probable that the main dipole electrification of the thundercloud is initiated as a consequence of freezing processes of one kind or another as soon as the ice-phase forms from super-cooled drops. The larger carriers of charge will then fall downwards and build up a field which may of itself be sufficient to cause spark break-down or may have to be enhanced by the selective charging mechanism of C. T. R. Wilson. Another method of enhancement, by droplets bouncing off melting hail in the lower portion of the cloud, has been discussed by Workman and Reynolds.[6]

45. The Charge on Rain from Electrified Clouds.

Although the predominant sign of the electric field below a thundercloud is negative (§ 33) it has long been known that the charge on the rain from it is mainly positive in sign. This can be qualitatively explained by C. T. R. Wilson's selective ion-capture mechanism (§ 44 (a)), which would operate as in Fig. 20 with all the signs in the diagram reversed and might be expected to be very effective since point-discharge at the ground creates an abundance of positive ions most of which will be converted by capture-nuclei of various kinds into heavy ions during their passage upwards to the cloud.

Simpson,[6] in an important series of observations made at Kew, has compared the current, i, carried down by rain to a square centimetre of the ground with simultaneous records of the rate of rainfall, R, and of the point-discharge

current, I, from an elevated point. For potential gradients exceeding 2000 v./m. he finds the relation $i \propto -I \times R^{0.57}$ from which it is clear that (i) the sign of the charge on the rain in these fields is always opposite to that of the prevailing field (mirror-image effect) ; (ii) with a given rate of rainfall the rain-current is directly proportional to the point-discharge current, and (iii) with a constant point-discharge current the rain-current varies with the rate of rainfall, increasing or decreasing approximately as $\sqrt{R}$.

It is clear from these results that the rain derives its charge from point-discharge ions. Ultimately, in very heavy rain, all the point-discharge ions are swept up and brought back again to earth, the rain current being found equal to the estimated point-discharge current per square centimetre of the ground. Whether the Wilson mechanism alone can account quantitatively for these results is still under discussion ; it is not impossible that more than one process is at work in charging the falling drops. The rain current by itself gives a statistical mean for the charge on the drops and there is evidence that individual drops can carry charges of opposite sign.

In light steady rain and in some showers, the electric field at the ground is found by Simpson to be generally negative and as high as -1000 v./m. When it is positive it seldom exceeds the fine weather gradient of about $+400$ v./m. Such rain is found to be positively charged, the charge on it being directly proportional to the departure of the field from its fine weather value. No explanation of this effect has as yet been suggested.

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SUBJECT INDEX

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